TITLE:

Collisional bending of the western Paleo-Kuril Arc deduced from paleomagnetic analysis and U–Pb age determination

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CITATION:

Katagiri, Takahiro ...[et al]. Collisional bending of the western Paleo-Kuril Arc deduced from paleomagnetic analysis and U–Pb age determination. *Island Arc* 2020, 29(1): e12329.

ISSUE DATE:

2020-01-22

URL:

<http://hdl.handle.net/2433/245933>

RIGHT:

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1 Original article

2 **Collisional bending of the western Paleo-Kuril Arc deduced**
3 **from paleomagnetic analysis and U-Pb age determination**

4

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13

14 **Abstract**

15 The Paleo-Kuril Arc in the eastern Hokkaido region of Japan, the westernmost part of the Kuril Arc in the northwestern
16 Pacific region, shows a tectonic bent structure. This has been interpreted, using paleomagnetic data, to be the result of
17 block rotations in the Paleo-Kuril Arc. To understand the timing and origin of this tectonic bent structure in the Paleo-
18 Kuril arc-trench system, paleomagnetic surveys and U-Pb radiometric dating were conducted in the Paleogene Urahoro
19 Group, which is distributed in the Shiranuka-hill region, eastern Hokkaido. The U-Pb radiometric dating indicated that
20 the Urahoro Group was deposited at approximately 39 Ma. Paleomagnetic analysis of the Urahoro Group suggested
21 that the Shiranuka-hill region experienced a 28° clockwise rotation with respect to East Asia. The degree of clockwise
22 rotation implied from the Urahoro Group is smaller than that of the underlying Lower Eocene Nemuro Group (62°) but
23 larger than that of the overlying Onbetsu Group (-9°). It is thus suggested that the Shiranuka-hill region experienced a
24 clockwise rotation of approximately 34° between the deposition of the Nemuro and Urahoro groups (50–39 Ma), and
25 a 38° clockwise rotation between the deposition of the Urahoro and Onbetsu groups (39–34 Ma). The origin of the
26 curved tectonic belt of the Paleo-Kuril Arc was previously explained by the opening of the Kuril Basin after 34 Ma.
27 The age constraint for the rotational motion of the Shiranuka-hill region in this study contradicts this hypothesis.
28 Consequently, it is suggested that the process of arc-arc collision induced the bent structure of the western Paleo-Kuril
29 Arc.

30

31 **Keywords**

32 Arc-Arc collision, Paleo-Kuril Arc, Paleomagnetism, U-Pb radiometric dating

33 1. Introduction

34 Bent structures of tectonic belts are globally common in various tectonic settings. They are recognized from
35 their contorted spatial distribution and rotation of terranes, as detected by paleomagnetic declination (Duermeijer, Nyst,
36 Meijer, Langereis & Spakman, 2000; Otofujii, Enami, Yokoyama, Kamiya, Kuma, Saito & Matsuda, 1999; Takahashi
37 & Saito, 1997). These bent structures of tectonic belts can be formed as a result of large-scale tectonic events, including
38 convergence of continental blocks or opening of back-arc basins, which generally accompany orogeny (e.g. Kano,
39 2002; Klootwijk, Conaghan, & Powell, 1985; McCabe, 1984). For example, the opening of the Japan Sea (the back arc
40 basin of Japanese Island arcs) and collision of the Izu-Bonin Arc caused a large-scale bent structure of the tectonic belts
41 of the Japanese Island arc system (e.g. Otofujii, 1996; Takahashi & Saito, 1997). Similarly, the collision of Gondwana
42 with Laurasia produced the curved mountain belts of the Cantabria-Asturias arc (Weil, Voo, & Pluijm, 2001).

43 Tectonic belts in the western part of the Paleo-Kuril Arc, the eastern part of the Hokkaido Island of
44 northernmost Japan, exhibit a large-scale bent structure (Fujiwara, Kanamatsu, & Nanayama, 1995; Kimura, 1990,
45 1993). Structural trends of the tectonic belts in this region significantly vary at the Kushiro marsh (Figure 1). Pre-
46 Neogene terranes distributed in the Konsen-coastal region extend from east to west, while they extend from north to
47 south in the Shiranuka-hill region (Figure 1). In addition, clockwise-deflected paleomagnetic declinations are reported
48 for those in the western region. Thus, tectonic belts in the Paleo-Kuril Arc originally extended in an east-west direction
49 parallel to the Arc-Trench system, and the tectonic belts in this area were bent after their deposition (Fujiwara et al.,
50 1995; Fujiwara & Kanamatsu, 1994; Kanamatsu, Nanayama, Iwata, & Fujiwara, 1992; Kimura, 1990, 1993; Nanayama,

51 Kanamatsu, & Fujiwara, 1993).

52 The formation of the bent structure in the Paleo-Kuril Arc occurred during the Eocene (Fujiwara et al., 1995;
53 Figure 2). Paleomagnetic surveys in the Shiranuka-hill region in the western part of the bent structure (Figure 1) provide
54 the age constraints for this tectonic event. The Upper Cretaceous to Middle Eocene Nemuro Group shows clockwise
55 deflections in paleomagnetic declination from the geographic north, while the uppermost Eocene to early Oligocene
56 Onbetsu Group shows no deflection (Fujiwara et al., 1995; Hamano, Tsunakaawa, Saito, & Kikawa, 1986; Kaiho, 1983,
57 1984). Although the paleomagnetic direction from the Urahoru Group, which is on the Nemuro Group and overridden
58 by the Onbetsu Group, was also measured by Fujiwara et al. (1995), it is not reliable because of an inaccurate process
59 in their demagnetization experiments (see later discussion). Therefore, the existing paleomagnetic data suggest that the
60 bent structure was formed between the depositions of the Nemuro and Onbetsu groups (Middle Eocene–Early
61 Oligocene).

62 The origin of the bent structure is ambiguous due to this weak constraint of the age of formation. Between the
63 deposition of the Nemuro and Onbetsu groups, there were two significant tectonic events (Figure 2). One was a collision
64 between the Paleo-Kuril Arc and Paleo-Northeastern Japan Arc and the other was a spreading of the Kuril Basin. The
65 former event was estimated to have started before the deposition of the Urahoru Group (Iijima, 1996; Kiminami, 2010;
66 Kimura & Kusunoki, 1997), although the latter was estimated to have occurred after the deposition of the Urahoru
67 Group (Fukusawa & Ishihara, 1992; Iijima & Tada, 1990; Kimura & Tamaki, 1985; Maeda, 1990).

68 In order to determine the age of formation of the bent structure precisely, we conducted paleomagnetic surveys

69 and U-Pb radiometric dating of the Urahoro Group. Our data implies that the bent structure of the western end of the
70 Paleo-Kuril Arc was formed both before and after the deposition of the Urahoro group. Because the block rotation
71 occurred before the deposition of the Urahoro Group, it cannot have been due to the spreading of the Kuril Basin and
72 therefore we conclude that the bent structure was mainly caused by the collision of the Paleo-Kuril Arc and the Paleo-
73 Northeastern Japan Arc.

74

75 2. Geological setting

76 Hokkaido Island is located in the northern part of the Japanese Archipelago. Tectonic belts in Hokkaido Island
77 were formed along the two arc-trench systems: the western Paleo-Northeastern Japan Arc and the eastern Paleo-Kuril
78 Arc, which collided with each other in the Paleogene (Kiminami, 1989, 2010; Kimura, 1994a; Kimura & Kusunoki,
79 1997). Hokkaido is composed of five tectonic belts: are the Oshima, Sorachi-Yezo, Hidaka, Tokoro and Nemuro belts
80 from the west to the east (Kiminami, 1986; Niida, 2010; Figure 1).

81 The Paleo-Northeastern Japan Arc was the subduction zone of the Pacific plate into the Eurasian plate. The
82 Oshima and Sorachi-Yezo belts that are distributed to the west of Hokkaido (Figure 1) were formed along the western
83 Paleo-Northeastern Japan Arc, and consist of the Jurassic to Cretaceous accretionary complexes and Cretaceous forearc
84 basin deposits (Kiminami, 1986; Kito, 1987; Takashima, Nishi, & Yoshida, 2006; Takashima, Yoshida, & Nishi, 2001).
85 These belts extend in a north-south direction and show eastward rejuvenation (Kiminami, 1986; Figure 1). On the other
86 hand, the Paleo-Kuril Arc was the subduction zone of the Kula or Izanagi plate to the Okhotsk plate (Kimura, 1994a;
87 Maeda, 1990; Figure 1). The Tokoro and Nemuro belts, located to the east of Hokkaido, were formed as a part of the
88 Paleo-Kuril Arc (Kiminami, 1986). They consist of the Upper Cretaceous to Paleogene accretionary prisms and the
89 Upper Cretaceous to Paleogene deposits (Kiminami, 1983; Naruse, 2003). The Hidaka Belt is located between the two
90 arcs, and contains accretionary complexes of both paleo-arcs (Komatsu, 1986; Toyoshima, Komatsu, & Shimura, 1994).
91 It consists of the Upper Cretaceous and Paleogene accretionary complex, and volcanic and metamorphic rocks
92 (Toyoshima et al., 1994).

93 The middle Eocene Urahoru Group is located in Kushiro, eastern Hokkaido, and belongs to the Nemuro
94 belt (Figures 1 and 2). It is 735–780 m thick, contains non-marine-to-littoral clastics, and is composed of conglomerate,
95 sandstone, siltstone and coal (Mabuti, 1962; Matsui, 1962; Sasa, 1940a, 1940b; Sato et al., 1968; Kawai, 1956). The
96 Urahoru Group unconformably overlies the Upper Cretaceous to middle Eocene Nemuro Group, and is unconformably
97 overlain by the upper Eocene to lower Oligocene Onbetsu Group (Mabuti, 1962; Matsui, 1962). The Urahoru Group is
98 composed of six Formations: The Beppo, Harutori, Tenneru, Yuubetsu, Shitakara and Shakubetsu Formations in
99 ascending stratigraphic order (Sasa, 1940a, 1940b). The Rushin Formation, which is correlated with the Beppo,
100 Harutori and Tenneru Formations, is observed at the westernmost part of the exposure (Tanai, 1957).

101 The Urahoru Group crops out in two regions: the eastern Konsen-coastal and the northwestern Shiranuka-hill
102 regions (Mabuti, 1962; Matsui, 1962; Sasa, 1940a, 1940b; Figures 1 and 2). The Quaternary deposits are located in the
103 Kushiro Marsh, which is located between these two regions, while the subsurface survey indicated that the Urahoru
104 Group continuously extends in the subsurface of the Kushiro Marsh (New Energy and Industrial Technology
105 Development Organization, 1990). Consequently, the Urahoru Group was formed in a single sedimentary basin (New
106 Energy and Industrial Technology Development Organization, 1990).

107 Geologic structures of the Urahoru Group, however, are significantly different between the two regions
108 (Figure 3). In the Konsen-coastal region, the bedding strikes are 140° north and the dips are 8° to the south. Tectonic
109 folds are not obvious in this region and NW-SE directed faults are commonly observed (Mabuti, 1962). In contrast, in
110 the Shiranuka-hill region, the NNE-SSW oriented tectonic folds and faults occur commonly, and faults that strike in

111 other directions (NW-SE, N-S and E-W) are also observed (Mabuti, 1962). Matsui (1962) suggested that some of these
112 folds and faults were formed synchronously with the deposition of the Urahoru Group.

113 The depositional age of the Urahoru Group is estimated to be the Bartonian (middle Eocene). Based on the
114 occurrence of macrofossils and benthonic foraminifers, Sasa (1940b) and Kaiho (1983) correlated the Urahoru Group
115 to the lower to middle part of the Poronai Formation of the Ishikari Group from which 36.7 Ma fission track age was
116 reported. Kimura and Tsuji (1990) demonstrated that the depositional age of the bentonite tuff bed intercalated in the
117 Rushin Formation is 38.0 Ma, based on a determination of the zircon fission-track-age. Recently, Katagiri, Naruse,
118 Hirata, and Hattori (2016) reported the depositional age of an acidic tuff bed intercalated in the Tenneru Formation as
119 (39.06 ±0.23) Ma by using the U-Pb radiometric dating method.

3. Methodology

3.1. Paleomagnetism

Samples for paleomagnetic analysis were collected from 28 sites in the Urahoro Group at the Shiranuka-hill region (Figures 2 and 3). The samples were taken by hand at 25 sites, while an electric-powered core drill was used at the remaining three sites. Five to eight blocks or core samples, which were oriented with a magnetic compass, were collected from each site. All samples were cut into cylindrical specimens, 25 mm in diameter and 22 mm in length.

Natural remanent magnetizations (NRM) were measured by using a superconducting rock magnetometer (model 760R of 2G Enterprises). One or two specimens were selected from each site for pilot analysis and were subjected to progressive thermal and/or alternating-field (AF) demagnetization experiments. The demagnetization experiments were conducted until the specimens showed erratic magnetic behavior. Because more stable magnetic behaviors of NRM were observed during the thermal demagnetization experiments, progressive thermal demagnetization experiments were conducted for the remaining specimens. The direction of a characteristic remanent magnetization (ChRM), which showed a linear trend of vector-endpoints decaying toward the origin of the orthogonal-endpoint diagram, was determined from each specimen by applying principal component analysis (Kirschvink, 1980) to the linear trend composed of at least three consecutive data points. Site mean directions of the ChRMs were calculated after Fisher's (1953) method at sites with four or more specimens showing ChRMs with 5.0° or less maximum angular deviation (MAD). In this study, site mean directions with 95% confidence interval (α_{95}) less than 20° were regarded as reliable records of the ancient geomagnetic field, and were used for calculating a group mean and determining paleomagnetic

polarities. Site mean directions with α_{95} larger than 20° were used only in determining its polarity.

In order to clarify magnetic mineralogy, strong-field thermomagnetic experiments with a horizontal magnetic balance and thermal demagnetization experiments of composite isothermal remanences (IRMs) were performed on selected samples. In the thermomagnetic experiments, small chips of sediment samples packed in quartz cups (5 mm in diameter and 10 mm in height) were heated to approximately 720°C and cooled in air in magnetic field of 0.25–0.6 T. The heating and cooling rates were approximately $8^\circ\text{C}/\text{min}$. Standard paleomagnetic specimens were used for the thermal demagnetization experiments of IRMs. Following the method of Lowrie (1990), IRMs were imparted along the orthogonal three axes of each specimen in magnetic fields of 2.5 T, 0.4 T, and 0.12 T by using a pulse magnetizer (Magnetic Measurements Ltd. MMPM-10). Stepwise thermal demagnetizations of IRMs were performed at temperatures up to 680°C in air. Remanent magnetization was measured with a 2G Enterprises superconducting magnetometer.

3.2. Zircon U-Pb radiometric dating

Samples for U-Pb radiometric dating were taken from two tuffaceous siltstone layers, named Ht1 and Ht2, respectively (Figures 2 and 3). We collected these samples from the subsurface outcrops along the tunnels of Kushiro Submarine Coal Mine (Figure 3). Ht1 and Ht2 are intercalated in the Harutori Formation, which are interpreted to be alluvial plain deposits (Sawaki et al., 2012). Ht1 is 1.25 m thick and consists of tuffaceous siltstone containing abundant coaly fragments. The sample was collected from the basal part of the bed. Ht2 is 22 cm thick and is composed of

156 tuffaceous clay.

157 The zircon grains for the analysis were mounted in a resin after being separated from the rock samples by using
158 the gravity concentration and magnetic separation techniques. In these procedures, the samples were crushed and grains
159 of over 250 μm in diameter were removed using a sieve. After the heavy minerals were concentrated by elutriation,
160 ferromagnetic minerals were then also removed. Heavy liquid separation was then performed by using a solution of
161 sodium polytungstate (specific gravity = 2.9). Finally, the zircon grains were handpicked and mounted on slide glasses
162 in epoxy resin (Specifix-20, Struers). Mounted grains were polished to expose the centers of individual grains.

163 Dating of zircon by U-Pb analysis was carried out using laser ablation-ICP-mass spectrometry (LA-ICPMS) with
164 a multiple collector-ICP-MS (Nu Instruments, Wrexham, UK) equipped with a 193 nm Excimer laser ablation system
165 (Analyte Excite 500, Teledyne Cetac, Omaha, USA) (Hattori, Sakata, Tanaka, Orihahi, & Hirata, 2017; Obayashi,
166 Tanaka, Hattori, Sakata & Hirata, 2017). The system setup and operational conditions are summarized in Table 1. An
167 ablation pit size of 20 μm was used throughout this study. Prior to the U-Pb isotopic analyses, both backscattered-
168 electron (BSE) and cathodo-luminescence (CL) images were created using a scanning electron microprobe (JEOL JXA-
169 8105). For zircon grains having clear zoning textures in the BSE or CL images, U-Pb dating was conducted on both
170 core and rim of a grain (two or three spots were selected), otherwise one spot was chosen for each grain. Mass bias
171 factors in the LA-ICPMS system were externally corrected by normalizing the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio for the NIST SRM610
172 (0.9096; Jochum & Stoll, 2008) and $^{206}\text{Pb}/^{238}\text{U}$ ratio for the Nancy 91500 (0.1792; Wiedenbeck et al., 1995). In addition,
173 the $^{206}\text{Pb}/^{238}\text{U}$ ratios for OD3 (^{238}U - ^{206}Pb age = 32.853 ± 0.016 Ma; Iwano et al., 2013; Lukács et al., 2015) were

174 measured as a secondary standard. Overall, the average ^{238}U - ^{206}Pb age of the OD3 was (32.72 ± 0.23) Ma ($n = 17$),
175 suggesting that the estimated ages agree with the literature values, within analytical uncertainties.

176 Concordia diagrams were made by using Isoplot 4.15 (Ludwig, 2012). A weighted mean value of ages of
177 measured grains was then calculated to estimate the eruptional/depositional age of each sample. Error ranges shown in
178 this manuscript are 95% confidence limits. Discordant data were excluded from the calculation of mean ages. When
179 two or more concordant data were obtained from a grain and there was no significant difference between the data, the
180 datum with the smallest error range was selected as the representative age of the grain. Error ranges of individual grains
181 were determined according to Sakata, Hattori, and Iwano (2014) and Katagiri et al. (2016).
182

4. Results

4.1. Paleomagnetic analysis

Initial NRM intensities of the specimens generally ranged from 10^{-10} Am² to 10^{-9} Am², whereas all specimens from three sites (1615, 1616, and 1621) and one specimen of Sites 1509 and 1611 had initial NRM intensities of approximately 10^{-8} Am². Low-stability components were generally observed in the demagnetization level below 280 °C (Figure 5 a-e). These components were regarded as viscous overprints of the recent geomagnetic field because they generally had northward and downward direction. ChRMs were isolated from specimens of nine sites after the removal of the low-stability components (Table 2). ChRMs were apparent in the range from 280 °C to 600 °C (Figure 5a-e and Table 2). In eight sites (1501, 1506, 1507, 1608, 1612, 1615, 1616, and 1617), four or more specimens showed ChRMs with $MAD \leq 5^\circ$ (Table 2). Samples from Sites 1615 and 1616 provided stable behaviors of ChRMs up to 600 °C (Figure 5a-e), while those of other sites showed erratic magnetic behaviors, that is, abrupt increase of remanent intensity and erratic directional change of NRM, at higher demagnetization levels above 360-480 °C (Figure 5c-d). Other specimens also showed erratic behaviors before producing ChRMs (Figure 5f).

Out of the above eight sites, six sites (Sites 1501, 1506, 1507, 1608, 1615, and 1616) yielded site-mean directions of ChRMs with $\alpha_{95} \leq 20^\circ$, while site means from Sites 1612 and 1617 had α_{95} exceeding 20° (Table 2). The site-mean directions of the remanent six sites exhibited an antipodal relationship with a NNE-SSW trend after tilt correction (Figure 7). The fold test based on McElhinny (1964) indicated that the ChRMs were acquired before the tilting. κ_a/κ_b (= 9.4) was larger than F_c (= 2.98), where κ_a and κ_b are concentration parameter of the overall mean direction after and

201 before tilt correction, respectively, and F_c is the F statistic, with degrees of freedom = 10 at 95% confidence limit. The
202 means of tilt-corrected site-mean directions from the six sites are: Declination (D) = 37.3°, Inclination (I) = 51.5°, α_{95}
203 = 13.9° and κ = 24.3, which is regarded as the paleomagnetic direction of the Urahoro Group.

204 Induced magnetization (Js) curves of samples with stable magnetic components at thermal demagnetization
205 levels up to 600 °C (Figure 6a-b) showed inflections at around 400 °C and 580 °C during heating in air, and decay to
206 about 620 °C (Figure 6a). The soft (< 0.12 T) and medium (0.12–0.4 T) components of composite IRMs exhibited a
207 large decrease of IRM below 400 °C and were demagnetized at 600 °C or 620 °C and the high components (0.4–2.5 T)
208 were demagnetized at 680 °C (Figure 6a).

209 Samples with erratic magnetic behaviors of NRM in high temperature demagnetization steps above
210 approximately 400 °C (Figure 5c-f) also exhibited inflections at around 400 °C and 580 °C for Js curves during heating
211 in air (Figure 6b, c). The three IRM components were demagnetized completely at 580 °C. An increase of Js above
212 550 °C was remarkable for samples with no stable component of NRM at thermal demagnetization temperatures
213 exceeding approximately 400 °C (Figure 6c).

214 Results of the thermomagnetic experiments and thermal demagnetization of composite IRMs suggested that
215 magnetite was a principal magnetic mineral. The observed inflection of Js curves at approximately 400 °C also indicated
216 the presence of maghemite or partially-maghematized magnetite. An inversion of maghemite to hematite during heating
217 in air occurred between 250 °C and 750 °C (Dunlop & Ozemir, 1997), which generated the decrease of Js and IRM
218 carried by maghemite (de Boer & Dekkers, 1996). Some samples exhibited Curie temperatures of 600–620 °C (Figure

6a). The Curie temperature above 600 °C has been reported for maghemite (Dunlop & Ozemir, 1997; Gehring et al., 2009). It is therefore suggested that the ChRMs isolated in this study are carried principally by magnetite and/or maghemite. Thermal demagnetization behaviors of the hard IRM components indicated the presence of hematite for some samples (Figure 6A). The increase of J_s above 500 °C on the thermomagnetic curve in air can be caused by the formation of magnetite, which is probably attributed to the thermal alteration of Fe-bearing paramagnetic minerals. Erratic magnetic behavior during thermal demagnetization in air, which makes it difficult to isolate stable magnetic components of NRM, was possibly related to the presence of such a newly formed magnetite.

4.2. U-Pb radiometric dating

Ht1: Measurements on 142 spots in 96 grains and concordant data were obtained from 71 grains (Table 3). Figure 8 shows a histogram of age distribution and a concordia diagram for all concordant data, except for the datum from grain S62, which shows an exceptionally old age (Table 3). The age data exhibited a unimodal distribution around 39 Ma with two outlier data (from grain S12 and S62) (Figure 8; Table 3). The main cluster of the distribution is composed of 69 concordant data with a weighted mean of (39.54 ± 0.17) Ma (Ludwig, 2012). The outlier data of grains S12 and S62 were 43 Ma and 5063 Ma, respectively. The outlier grains were different from other grains in roundness of grain outlines and colors in the CL image (Figure 9).

Ht2: Measurements were conducted on 74 spots in 58 grains; concordant data were recorded from 36 grains (Figure 8). The age data were clustered into two groups with five outlier data. The main cluster was composed of 27

237 data with a weighted mean of (56.18 ± 0.42) Ma. The small cluster was composed of 4 data with a weighted mean of

238 (40.8 ± 1.1) Ma. Other five outlier data were 63 Ma, 92 Ma, 252 Ma, 1746 Ma, and 1880 Ma (Figure 8; Table 4). Grains

239 showing very old ages (1746 Ma and 1880 Ma) were clearly different from the other grains in roundness of grain

240 outlines and colors in CL images (Figure 10).

241

242 5. Discussion

243 5.1. Depositional age of the Urahoru Group

244 Radiometric dating conducted in this study indicated that tuffaceous layers in the Urahoru Group were
245 deposited at about 40 Ma. The calculated depositional age of Ht1 and Ht2 were (39.54 ± 0.17) Ma and (40.8 ± 1.1) Ma,
246 respectively, and these are concordant to the depositional age of the tuff layer Tn1 (39.06 ± 0.23 Ma) in the Tenneru
247 Formation (Figure 4; Katagiri et al., 2016). The depositional ages of the layers were calculated as the weighted mean
248 values of the radiometric ages of zircons except for the grains that were considered to be detrital in origin. The grains
249 S62 and S12 of the Ht1 tuffaceous layer were interpreted to be of detrital origin, due to their old radiometric ages and
250 abraded morphology (Figure 9; Table 3). Multimodal distribution of radiometric ages of zircon grains in the Ht2
251 tuffaceous layer implies that Ht2 also contained abundant zircon grains of detrital origin (Figure 8; Table 4). This
252 interpretation was supported by variations in colors, morphologies and sizes of individual grains in CL images (Figure
253 10). Grains of detrital origin were excluded from the age calculation and only the grains of the younger cluster were
254 used.

255 Stratigraphic relationships among the tuffaceous layers with Tn1 and layers with magnetic polarities are
256 shown in Figure 4. The normal polarities from Sites 1501, 1506, 1507, 1608, 1612, 1616, and 1617 were correlated to
257 the polarity chron C18n.1n, ranging from 39.6 Ma to 38.6 Ma (Ogg, 2012: Figures 3 and 11) because the Tn1 layer
258 with an age of (39.06 ± 0.23) Ma was stratigraphically intercalated in the intervals of these sites (Katagiri et al., 2016;
259 Ogg, 2012; Figures 4 and 11). The lower reversed polarities of Site 1615 were correlated with C19r, C18r, or C18n.1r

260 based on the ages of Ht1 (39.54 ± 0.27) Ma and Ht2 (40.8 ± 1.1) Ma. Thus, we suggest that the deposition of the Urahoru
261 Group started before 39.6 Ma, which is the beginning of C18n.1n. Our depositional age estimation of the Urahoru
262 Group does not contradict previous estimates (Kaiho, 1983; Kimura & Tsuji, 1990), but gives a much stronger constraint
263 to the age of formation of the Urahoru Group.

264

265 *5.2. Paleomagnetic direction of the Urahoru Group*

266 The paleomagnetic direction of the Urahoru Group in the Shiranuka-hill region derived in this study was
267 characterized by a clockwise deflected declination (Table 5). Compared with the direction of the geocentric axial dipole
268 field (GADF) expected at the Shiranuka-hill region ($D = 0^\circ$ and $I = 57.5^\circ$), the deflection in declination was significant
269 ($37.3 \pm 22.7^\circ$) while that in inclination was not significant ($6.0 \pm 13.9^\circ$). Fujiwara and et al. (1995) also reported a
270 clockwise-deflected paleomagnetic direction from the Urahoru Group in the Shiranuka-hill region (Table 5). While the
271 inclination value coincides with that of this study, the declination of Fujiwara et al. (1995) is larger than that of our
272 study (Table 5). Because they conducted progressive demagnetization only for pilot specimens and only one treatment
273 temperature was applied for each site (Fujiwara et al., 1995), this discrepancy is probably due to the effect of the low
274 stability components that were not removed sufficiently by Fujiwara et al. (1995). The directional data in Fujiwara et
275 al. (1995) were obtained by one-step heating of thermal demagnetization at 100 °C, 150 °C, or 300 °C. Our
276 demagnetization results indicate that their demagnetization levels, especially at 100 °C and 150 °C, were too low to
277 remove the low-stability components because ChRMs were isolated at demagnetization levels above 280 °C or more
278 in our study (Figure 5 and Table 2). In addition, Fujiwara et al. (1995) did not get data with reversed polarity, and did

279 not describe any demagnetization behaviors. Thus, we consider that our results are more reliable.

280 We compared the paleomagnetic direction of the Urahoro Group with an apparent polar wandering path
281 (APWP) for East Asia observed by Cogné et al. (2013) in order to examine tectonic movements of terranes in the Paleo-
282 Kuril Arc relative to East Asia. Comparing with an expected direction at the Shiranuka-hill region calculated from a 40
283 Ma paleopole of the APWP, our data indicate a clockwise rotation of $(28.0 \pm 23.3)^\circ$ and no significant northward
284 translation (Table 5). It is therefore suggested that the Shiranuka-hill region was subjected to a clockwise rotation of
285 about 29° after the formation of the Urahoro Group (*ca.* 39 Ma).

286

287 5.3. Rotational movement in the Shiranuka-hill region inferred from paleomagnetic data

288 Two previous studies in the Shiranuka-hill region, Fujiwara et al. (1995) and Hamano et al. (1986), reported
289 paleomagnetic directions of the Cretaceous to early Eocene Nemuro Group underlying the Urahoro Group, and those
290 directions exhibited a large clockwise deflection of about 70° in declination (Table 5). Here we adopted the data from
291 Hamano et al. (1986) as a paleomagnetic direction of the Nemuro Group because the data in Fujiwara et al. (1995)
292 contained very large errors (α_{95} was 88° in maximum). The paleomagnetic samples of Hamano et al. (1986) were
293 taken from the horizons with K-Ar and fission-track ages of 54 Ma, 52 Ma, and 47.7 Ma (Fujiwara et al., 1995; Kimura
294 and Tsuji, 1990; Shibata et al., 1984). Compared with a 50 Ma paleopole of the APWP for East Asia (Cogné et al.,
295 2013), their data indicated a clockwise rotation of about 62° for the Shiranuka-hill region after the deposition of the
296 Nemuro Group. The clockwise deflected declination of the Urahoro Group was smaller than that of the Nemuro Group

297 (Table 5): a discrepancy between the two declinations was $(33.9 \pm 26.4)^\circ$. This implies that the Shiranuka-hill region
298 experienced a regional clockwise rotation of about 34° between the depositions of the Nemuro and Urahoro Groups
299 (ca. 50 Ma to 39 Ma: Figure 12).

300 Fujiwara et al. (1995) also reported a paleomagnetic direction of the Onbetsu Group overlying the Urahoro
301 Group, which exhibited no discrepancy with respect to East Asia, both in its declination and in inclination (Table 5).
302 The paleomagnetic samples of the Onbetsu Group were collected from the lower part of the Onbetsu Group (Fujiwara
303 et al., 1995), the depositional age of which was estimated to be the latest Eocene to early Oligocene, based on the
304 occurrence of the benthonic foraminifera (Kaiho, 1983; 1984). According to Fujiwara & Kanamatsu (1994), the
305 horizons sampled by Fujiwara et al. (1995) were correlated to the polarity chron C13n.r, which ranged from 35.0 to
306 33.7 Ma (Ogg, 2012). The paleomagnetic data from the Onbetsu Group (Table 5) infers no rotational motion of the
307 Shiranuka-hill region after the deposition of the Onbetsu Group (ca. 34 Ma). This implies that the clockwise rotational
308 motion of the Shiranuka-hill region after the formation of the Urahoro Group ceased at about 34 Ma (Figure 12). It is
309 noteworthy that the data from the Onbetsu Group by Fujiwara et al. (1995) have a large value of α_{95} , and the discrepancy
310 between the paleomagnetic directions of the Onbetsu and Urahoro Groups is not significant $(36.6 \pm 51.1)^\circ$.

311

312 5.4. Bent structure in the Paleo-Kuril Arc

313 Paleomagnetic declinations of the Paleo-Kuril Arc are summarized in Figure 13. All terranes located west
314 of the Kushiro marsh region exhibited clockwise rotated paleomagnetic directions, although the degree of rotation

315 differed significantly among individual terranes. In addition to the Shiranuka-hill region, paleomagnetic directions
316 indicating larger clockwise rotations were reported from the Tokoro and Hidaka belts, the westernmost part of the
317 Paleo-Kuril Arc (Fujiwara et al., 1995; Figure 13, Table 5). In contrast, paleomagnetic directions of the Nemuro Group
318 from the eastern part of the Paleo-Kuril Arc, namely the Konsen-coastal region, suggest that tectonic blocks east of the
319 Kushiro marsh region did not experience rotation after the formation of the Nemuro Group (Fujiwara & Kanamatsu,
320 1990; Nifuku, Kodama, Shigeta & Naruse, 2009; Tanaka & Uchimura, 1989; Figure 13, Table 5). These paleomagnetic
321 data imply that the western wing of the curved tectonic belt consist of some clockwise rotated terranes while the eastern
322 part is a non-rotated block (Figure 13). The boundary between the western and eastern block was inferred to be in the
323 Kushiro marsh region.

324 Geophysical data of the Paleo-Kuril Arc also imply that a large block boundary of the Paleo-Kuril Arc was
325 located in the Kushiro mash region (Figure 13). The gravity survey results of Yamamoto and Matsushima (1990) reveal
326 that a positive gravity anomaly, corresponding to the distribution of the Nemuro and Urahoro Groups, extends east to
327 west in the Konsen-coastal region, while the trend of the anomaly turns at the Kushiro marsh region and extends north
328 to south in the Shiranuka-hill region. The Urahoro Magnetic Anomaly Belt, which corresponds to the distribution of
329 ultramafic rocks in the Nemuro Group, was also bent offshore Kushiro (Ogawa & Sunakawa, 1976; Figure 13). These
330 geophysical data indicate that the entire tectonic block distributed west of the Kushiro marsh region experienced a
331 clockwise rotation or northward migration relative to the Konsen-coastal region, resulting in the bent structure of the
332 Paleo-Kuril Arc.

333

334 *5.5. Arc-Arc collision as a cause of the bent structure in the Paleo-Kuril Arc terranes*

335 This study suggests a clockwise rotation of 29° for the Shiranuka-hill region after the deposition of the
336 Urahoru Group (*ca.* 39 Ma). This implies that the clockwise rotational motion was initiated before the deposition of the
337 Urahoru Group (*ca.* 50 Ma), and a clockwise rotation of about 34° occurred between 50 Ma and 39 Ma. The rotational
338 motion in the Shiranuka-hill region possibly ceased before the formation of the Onbetsu Group (*ca.* 34 Ma).

339 The timing of the rotational motion for the Shiranuka-hill region between 50 Ma and 39 Ma suggests that
340 the block rotation occurred due to the arc-arc collision between the Paleo-Kuril Arc and the Paleo-Northeastern Japan
341 Arc. Previous studies have suggested that the region rotated due to the opening of the back-arc basin (the Kuril Basin)
342 of the Kuril Arc (Fujiwara & Kanamatsu, 1994; Fujiwara et al., 1995). However, the rotational motion in the region
343 before 39 Ma contradicts this hypothesis because the oldest estimate for the initiation of spreading of the Kuril Basin
344 was during the Oligocene, 34 Ma. Iijima and Tada (1990) and Fukusawa and Ishihara (1992) concluded that the Kuril
345 Basin was formed synchronously with the deposition of the Takkobu Formation, during the upper Oligocene to the
346 lower Miocene. Maeda (1990) suggested that the Kuril basin opened just after 16 Ma based on the cessation of the
347 Hidaka metamorphism and the coeval north-south trending extension. Kimura and Tamaki (1985) estimated that the
348 Kuril Basin was formed during the Oligocene to middle Miocene by comparing the basal depth and heat flow of the
349 Kuril Basin with those of the Shikoku Basin and South China Sea. Accordingly, the rotation of the Shiranuka-hill region
350 before 39 Ma cannot be attributed to the spreading of the Kuril Basin.

351 The rotation age estimated by this study is concordant with the timing of the initiation of the arc-arc collision,
352 which has been deduced from the transition of sedimentary basins (Iijima, 1996; Kiminami, 2010; Kimura & Kusunoki,
353 1997). In Hokkaido Island, the Paleogene strata, including the Urahoro Group, are characterized by coal-bearing sub-
354 aerial deposits with gravels supplied from the trenchside (Iijima, 1996; Nagahama, Terui, Nagahama, & Sato, 1978).
355 These features are in contrast to those of the underlying Upper Cretaceous forearc deposits, including the Nemuro
356 Group, which mainly consist of fine-grained marine deposits supplied from each paleo-arc zones (Naruse, 2003;
357 Takashima et al., 2001; Takashima, Kawabe, Nishi, Moriya, Wani & Ando, 2004). It has been suggested that the arc-
358 arc collision occurred just before the late Eocene on the basis of drastic changes in sedimentological features in the
359 basins, as well as the cessation of magmatism along the eastern margin of the Eurasian plate (Iijima, 1996; Kiminami,
360 2010; Kimura & Kusunoki, 1997). Kimura and Kusunoki (1997) suggested that the collision occurred before the back-
361 arc spreading, and our result supports this view. Therefore, we suggest that the Shiranuka-hill region was rotated due
362 to the oblique arc-arc collision of the Paleo-Kuril Arc and Paleo-Northeastern Japan Arc, which is a common cause of
363 block rotations of terranes (e. g. Kano, 2002; Takahashi & Saito, 1997).

364 **6. Conclusions**

365 Radiometric dating, using U-Pb, of zircons extracted from tuffaceous beds, and paleomagnetic analysis in the

366 Urahoru Group, were conducted in this study. Based on the U-Pb ages, the depositional age of the Urahoru Group was

367 estimated to be approximately 39 Ma. A paleomagnetic direction of the Urahoru Group in the Shiranuka-hill region

368 indicated a 29° clockwise rotation of the region after its deposition (*ca.* 39 Ma). In conjunction with previous

369 paleomagnetic data, our data indicated that the Shiranuka-hill region experienced an approximately 34° clockwise

370 rotation between the deposition of the Nemuro and Urahoru Groups (50–39 Ma) and 37° clockwise rotation between

371 the deposition of the Urahoru and Onbetsu Groups (39–34 Ma). The clockwise rotational motion of the Shiranuka-hill

372 region is probably representative of the block rotation of the western part of the Paleo-Kuril Arc, resulting in the

373 formation of the bent structure of the arc in the eastern Hokkaido Island. The age constraints of the block rotation

374 revealed in this study contradict the conventional hypothesis that the block rotation of the Paleo-Kuril Arc was formed

375 by the opening of the Kuril basin. Rather, our study supports the hypothesis that the collision between the Paleo-Kuril

376 Arc and Paleo-Northeastern Japan Arc caused the rotation of the tectonic blocks in the Paleo-Kuril Arc.

377 Acknowledgements

378 Measurements of paleomagnetism and radiometric ages were conducted at Kyoto University. We wish to thank
379 Harutaka Sakai, Tetsu Kogiso, and Tetsuo Kawakami who kindly let us use laboratory equipment for the radiometric
380 dating. Thanks are also due to Kota Takatsuka and Hideyuki Obayashi for their support in using the JOEL JXA-8105
381 superprobe, and ICP-MS, respectively. We also thank Naoyoshi Hasegawa of Shiranuka village, Takaori Ishikawa of
382 the Kushiro museum, Katsumi Uchida and Yoshihiro Ito of the Kushiro Coal Mine and all the members of the Akkesi
383 Marine Station for supporting us during the field survey in Kushiro. We appreciate helpful reviews by Tetsuji Muto,
384 Gaku Kimura and Koji Uno. We would like to thank Serem Gideon Kiptoo of INPEX and Editage (www.editage.com)
385 for English language editing. This work was supported by JSPS KAKENHI Grant Numbers JP26610161 and
386 JP26287127.

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568 **Figure Captions**

569 Figure 1. Geographic map of the Northwestern Pacific Region (a) and Geotectonic division of Hokkaido Island (b),
570 modified after Nakanishi et al. (2009) and Nida (2010). Hokkaido is located at the convergent boundary and consists
571 of western Paleo-Northeastern Japan Arc and eastern Paleo-Kuril Arc.

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573 Figure 2. Conventional tectonic model of Hokkaido based on Fujiwara and Kanamatsu (1994), Kiminami (2010,
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575 with each other. (c) The spreading of the Kuril basin caused a transpressional tectonic condition at the boundary of
576 the two arcs, resulting in the clockwise rotation of the Shiranuka-hill region. (d) Hidaka Orogeny occurred due to
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583 Figure 4. Stratigraphic relationships of horizons of the radiometric and paleomagnetic samples. Geological columns
584 of the western region were made from a route map along the Satombetsu river section (Figure 3d). Columns of the
585 eastern region were made from Miyasaka and Hoyanagi (2010). Black and white circles indicate normal and reversed

586 paleomagnetic polarities, respectively.

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596 after tilt corrections. Solid and open symbols are on the lower and upper hemispheres, respectively. Ovals around the

597 directions are 95% confidence limits. Numerals indicate site numbers.

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609 Figure 10. CL images of representative zircon grains obtained from Ht2. Measured points are shown with their
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- 622 Fm., Formation. G., Group.
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- 624 Table 1. Instrumentation and operational settings of the LA-ICP-MS technique for U-Pb dating.
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- 626 Footnotes: Lat, latitude (N°); Long, longitude (E°); n_A , number of specimens used in calculation of the site mean
- 627 direction; n , number of samples collected from a site. D_{IS} , I_{IS} : *in-situ* declination and inclination, respectively; D_{TC} ,
- 628 I_{TC} , declination and inclination, respectively, after tilt correction and true north correction; N, normal polarity; R,
- 629 reversed polarity; —, No data.
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- 633 Data with circles in the left column were used for calculation of the mean.
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- 635 Table 4. Calculated $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages of zircon grains from Ht2.
- 636 Footnotes: 7/5 Age, $^{207}\text{Pb}/^{235}\text{U}$ age (Ma); 6/8 Age, $^{206}\text{Pb}/^{238}\text{U}$ age (Ma). Uncertainties are reported at the 2σ level.
- 637 Data with circles in the left column were used calculation of the mean.
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- 639 Table 5. Paleomagnetic directions obtained from the Paleo-Kuril Arc.

640 Footnotes: D and I, tilt-corrected declination and inclination of the observed mean directions, respectively; ΔD and
641 ΔI , 95% confidence limit of declination and inclination, respectively. $\Delta D = \sin^{-1}(\sin(\alpha 95)/\cos(I))$, $\Delta I = \alpha 95$; n,
642 number of data points used in calculating mean directions; R and F, discrepancies in declination (R) and inclination
643 (F) between the observed and expected directions; ΔR and ΔF , 95% confidence limits of R and F, respectively.
644 Tectonic parameters were calculated based on the definition of Beck et al. (1980); —, no data; R. P., reference pole
645 of East Asia from Cogné et al. (2013). Paleomagnetic directions and paleomagnetic pole positions of the previous
646 studies, except for Hamano et al. (1986), were recalculated from site-mean directions with $\alpha 95 \leq 20$. The positions
647 of sampling sites A-E in Fujiwara and Kanamatsu (1990) were assumed as follows: A, 43.05°N, 144.84°E; B,
648 43.14°N, 145.18°E; C, 43.15°N, 145.16°E; D, 43.19°N, 145.53°E; E, 43.15°N, 145.23°E.
649

図 1.3

北海道西部，北海道中央部，北海道東部の田・紀藤 (1986) を一部修正。1=ジュラ紀付加体 (白亜紀前期)，2=白亜紀付加体I (白亜紀前期)，3=白亜紀前弧海盆堆積体 (蝦夷層群)，4=空知層群上部・空知層群下部，5=蛇紋岩，6=白亜紀付加体I (白亜紀前期)，7=白亜紀付加体II (白亜紀前期)，8=白亜紀付加体III (白亜紀後期)，9=白亜紀後期～古第三紀付加体 (日高層群・中の川層群)，10=古第三紀付加体 (湧別層群・広尾コンプレックス)，11=海山付加体 (仁頃層群)，12=白亜紀後期～古第三紀前弧海盆堆積体 (佐呂間層群・根室層群)。

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Fm., Formation. G., Group.

Table 1. Instrumentation and operational settings of the LA-ICP-MS technique for U-Pb dating.

Laser ablation system	Model	ESI NWR 193 (New Wave Research, Oregon, USA)
	Wavelength	193 nm
	Fluence	2.0 J/cm ² (40%)
	Frequency	4 Hz
	Spot size	20 μm
	Pre Abration	1 shot (for cleaning)
	Carrier gas	He
	He gas flow	0.45 l/min
	Ar-make up gas flow	0.94 l/min
	Number of Shots	50 shots/spit
ICP-MS	Model	Nu Plasma II MC-ICP-MS (Nu Instruments, Wrexham, UK)
	Analyzed isotopes	²⁰² Hg, ²⁰⁴ (Pb+Hg), ²⁰⁶ Pb, ²⁰⁷ Pb, ²⁰⁸ Pb, ²³² Th (Faraday), ²³⁸ U
	Integration time	12 s
Data analysis	Gas blank	Before, after and between individual measurement
	Primary standard	Nancy 91500 (²⁰⁶ Pb / ²³⁸ U) NIST SRM610 (²⁰⁷ Pb / ²⁰⁶ Pb)
	Secondary standard	OD3

Table 2. Paleomagnetic data from the Urahoro Group.

Site	Lat	Long	Lithology	n_A/n	Levels	D _{IS}	I _{IS}	D _{TC}	I _{TC}	α_{95}	κ	Polarity
1501	43.24	144.07	muddy sand	5/7	280–440	-3.5	55.9	24.5	53.3	16.7	18.7	N
1505	43.21	144.01	muddy sand	0/5	—	—	—	—	—	—	—	—
1506	43.23	144.00	muddy sand	6/8	280–480	56.0	44.3	23.4	49.2	10.0	45.5	N
1507	43.23	144.00	muddy sand	5/8	280–400	68.2	45.1	38.4	53.3	19.7	16.0	N
1508	43.23	144.00	muddy sand	0/5	—	—	—	—	—	—	—	—
1509	43.20	144.00	muddy sand	0/8	—	—	—	—	—	—	—	—
1512	43.24	144.07	muddy sand (nodule)	0/8	—	—	—	—	—	—	—	—
1601	43.23	144.07	muddy sand	0/6	—	—	—	—	—	—	—	—
1602	43.23	144.08	muddy sand	0/7	—	—	—	—	—	—	—	—
1603	43.23	144.08	muddy sand	0/8	—	—	—	—	—	—	—	—
1604	43.09	143.82	muddy sand	0/8	—	—	—	—	—	—	—	—
1605	43.09	143.82	muddy sand	0/7	—	—	—	—	—	—	—	—
1606	43.09	143.82	muddy sand	0/8	—	—	—	—	—	—	—	—
1607	43.09	143.83	muddy sand (nodule)	0/7	—	—	—	—	—	—	—	—
1608	43.09	143.82	muddy sand	6/8	280–320	-32.2	26.7	35.3	34.4	7.5	55.6	N
1609	43.09	143.82	muddy sand	0/7	—	—	—	—	—	—	—	—
1611	43.12	143.95	muddy sand	1/5	460–520	89.2	54.9	35.9	66.5	—	—	—
1612	43.12	143.95	muddy sand	4/8	280–320	35.9	40.2	23.9	1.3	31.4	9.5	N
1613	43.11	143.97	very-fine-grained sand	0/7	—	—	—	—	—	—	—	—
1615	43.09	143.82	muddy sand	8/8	280–600	117.4	-11.1	- 145.8	-73.9	7.8	50.9	R
1616	43.09	143.82	muddy sand	6/6	280–580	-42.7	42.7	62.1	40.3	5.8	134.0	N
1617	43.09	143.82	muddy sand	4/7	280–320	-24.0	34.2	43.8	32.1	40.9	10.2	N
1621	43.09	143.82	muddy sand	0/7	—	—	—	—	—	—	—	—
1622	43.11	143.97	very-fine-grained sand	0/8	—	—	—	—	—	—	—	—
1623	43.10	143.83	muddy sand	0/5	—	—	—	—	—	—	—	—
1627	43.09	143.83	muddy sand	0/7	—	—	—	—	—	—	—	—
1628	43.09	143.83	muddy sand (nodule)	0/7	—	—	—	—	—	—	—	—
1629	43.23	144.07	fine-grained	0/5	—	—	—	—	—	—	—	—
Overall mean	[6]			In situ				-12.0	49.5			
				Unfolded				37.3	51.5	13.9	24.3	

Footnotes: Lat, latitude (N°); Long, longitude (E°); n_A , number of specimens used in calculation of the site mean direction; n , number of samples collected from a site; D_{IS}, I_{IS}, in-situ declination and inclination, respectively; D_{TC}, I_{TC}, declination and inclination, respectively, after tilt correction and true north correction; N, normal polarity; R, reversed polarity; —, No data.

Table 3. Calculated $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages of zircon grains from Ht1.

Grain No		7/5Age	error	6/8Age	error
B1		48.85	1.59	39.32	0.59
B1	○	40.07	0.80	39.36	0.52
B2		46.05	0.81	40.35	0.52
B3	○	43.06	1.80	41.17	0.65
B3		44.71	2.26	40.05	0.69
B4	○	41.89	2.51	39.81	0.72
B4		52.11	1.42	40.64	0.59
B5		155.30	3.04	45.82	0.67
B5	○	40.23	0.93	39.84	0.54
B5		39.51	1.75	40.38	0.64
B6		39.62	0.89	38.80	0.52
B6		41.52	1.47	40.13	0.60
B6	○	39.41	0.92	39.37	0.53
B8	○	38.53	0.98	38.73	0.42
B8		39.58	0.68	38.56	0.37
B11	○	40.37	1.50	40.53	0.51
B12		39.29	0.96	38.70	0.42
B12	○	39.57	0.78	39.26	0.40
B12		41.62	1.34	40.34	0.48
B13	○	39.01	1.10	38.94	0.44
B14		40.39	0.68	39.23	0.38
B16	○	41.15	1.78	38.99	0.54
B17		42.74	2.36	39.60	0.62
B17		39.88	0.62	39.10	0.37
B18	○	40.18	1.72	40.03	0.54
B19	○	39.86	1.19	38.77	0.45
B19		37.68	2.12	39.03	0.64
B23		44.86	2.47	39.23	0.67
B24		41.20	0.68	39.37	0.45
B25	○	40.19	0.93	39.66	0.49
B25		40.10	1.47	39.85	0.56
B25		41.77	1.08	38.71	0.49
B26	○	39.67	1.37	40.04	0.55
B27	○	40.00	0.95	39.96	0.49
B27		41.30	1.10	39.96	0.51
B28		41.69	1.13	39.02	0.50
B28	○	39.63	1.26	39.21	0.53
B28		39.14	0.79	38.77	0.46
B30		782.42	10.98	102.40	1.66
B30	○	39.44	1.07	38.99	0.92
B31	○	40.55	1.68	39.16	0.97
B31		40.36	2.02	38.80	1.00
B32		74.20	2.33	42.02	1.03
B34		50.13	3.75	39.69	1.15
B34	○	40.58	1.52	39.45	0.97
B35		41.83	1.20	41.19	0.98
B35	○	39.87	1.05	39.38	0.92
B36	○	40.63	1.47	38.92	0.95
B37		924.95	20.96	121.17	3.59

B37 ○	39.61	1.19	38.71	0.92
B38 ○	39.05	1.18	38.58	0.92
B39	42.53	2.14	39.09	1.01
B40	40.55	1.43	39.91	0.69
B40 ○	40.02	0.99	39.50	0.64
B40	40.63	1.25	39.85	0.67
B41	39.29	1.01	39.82	0.65
B41 ○	39.87	0.94	39.46	0.63
B42 ○	41.13	2.77	39.25	0.83
B43	45.97	2.43	39.60	0.79
B44 ○	40.24	1.13	39.72	0.66
B45 ○	38.03	1.44	38.60	0.68
B46	40.17	1.26	39.82	0.67
B46 ○	40.13	0.81	40.04	0.63
B47 ○	39.61	1.12	38.66	0.64
B48	42.72	1.57	40.31	0.71
B49 ○	38.65	1.45	40.07	0.51
B51 ○	40.05	0.80	39.00	0.40
B51	39.58	2.13	38.70	0.59
B52 ○	39.30	1.45	39.61	0.50
B52	36.74	1.76	38.58	0.54
B53	39.05	1.22	39.95	0.47
B53	40.12	0.63	39.30	0.38
S1	90.91	3.38	44.20	0.98
S1	41.52	0.90	40.04	0.74
S2	41.55	0.93	40.39	0.75
S3	70.03	1.86	46.59	0.91
S4 ○	39.73	1.37	40.37	0.80
S5	45.35	2.26	39.40	0.87
S6	43.45	1.33	40.51	0.79
S7 ○	39.85	0.85	39.45	0.73
S8	41.41	1.33	40.01	0.79
S9 ○	40.78	1.80	40.14	0.84
S10	40.09	0.99	40.06	0.76
S10 ○	40.61	0.92	40.89	0.76
S11 ○	40.38	0.94	40.24	0.75
S12 ○	44.12	0.94	43.43	0.53
S13 ○	40.45	0.91	40.18	0.50
S14 ○	39.32	0.78	38.93	0.47
S15 ○	40.48	1.78	40.05	0.62
S16	46.41	1.46	39.11	0.55
S17	453.38	7.36	96.99	1.47
S17	42.99	1.11	40.07	0.52
S18 ○	40.50	1.66	40.27	0.61
S19 ○	39.99	1.11	38.72	0.51
S21 ○	39.52	1.17	39.45	0.53
S22	41.06	1.07	39.62	0.51
S22 ○	40.44	0.65	40.34	0.47
S23	47.17	1.12	42.18	0.54
S24	615.26	10.57	82.38	1.54
S25	171.74	2.54	46.84	0.69
S26	38.41	1.76	39.03	0.70
S26 ○	39.59	1.46	38.59	0.65
S27 ○	40.19	0.78	39.72	0.58

S28 ○	38.25	0.69	37.77	0.55
S28	47.94	0.85	40.75	0.60
S29 ○	40.12	0.80	40.32	0.60
S30 ○	40.42	1.47	40.01	0.67
S31	42.51	0.97	41.35	0.63
S32 ○	38.97	1.02	39.61	0.61
S33 ○	39.14	1.21	38.09	0.61
S34 ○	40.50	1.23	39.52	0.63
S35 ○	39.49	1.14	38.96	0.88
S36	41.46	1.29	40.22	0.92
S37 ○	42.04	1.86	41.33	1.00
S38 ○	42.34	1.30	42.04	0.96
S39 ○	40.11	1.02	39.46	0.88
S40	40.73	1.04	39.64	0.88
S41	42.50	1.01	39.63	0.88
S42	41.40	1.07	40.02	0.89
S43 ○	39.98	1.47	39.74	0.93
S44 ○	39.69	1.23	39.28	0.90
S45 ○	40.36	1.08	39.54	0.89
S45 ○	41.12	1.48	41.47	0.97
S46	40.77	0.99	39.82	0.88
S47	40.83	1.02	39.62	0.86
S48 ○	40.16	1.06	39.59	0.86
S49	40.66	1.28	39.28	0.87
S50 ○	41.80	1.68	41.12	0.95
S51 ○	40.03	1.53	39.71	0.91
S52 ○	40.75	1.43	39.77	0.90
S52	42.11	1.67	39.51	0.92
S53 ○	39.74	1.01	40.02	0.87
S54 ○	39.91	1.74	39.52	0.93
S54	48.67	1.75	40.81	0.94
S55 ○	40.24	0.98	40.15	0.87
S56 ○	40.88	1.17	40.22	0.88
S57	50.06	1.24	40.51	0.88
S58	41.61	1.19	39.06	0.71
S59	40.28	0.83	39.55	0.69
S60	40.79	0.86	39.99	0.70
S61	68.81	1.66	42.69	0.77
S62	5018.62	26.54	5063.27	91.84

Footnotes: 7/5 Age, $^{207}\text{Pb}/^{235}\text{U}$ age (Ma); 6/8 Age, $^{206}\text{Pb}/^{238}\text{U}$ age (Ma). Uncertainties are reported at the 2σ level. Data with circles in the left column were used in mean calculation of the mean.

Table 4. Calculated $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages of zircon grains from Ht2.

Grain No	7/5Age	error	6/8Age	error
B1	1803.77	39.05	613.09	23.76
B3	428.15	4.26	299.80	2.70
B4	198.69	1.84	157.49	1.24
B4	153.64	1.38	151.68	1.16
B5	55.49	0.82	54.89	0.46
B6	57.59	0.96	56.43	0.49
B7	130.81	1.23	72.23	0.56
B7	59.93	0.67	57.16	0.45
B8	56.49	0.68	56.23	0.44
B10	1503.94	6.55	446.30	3.36
B10	256.53	2.20	212.42	1.65
B11	78.56	1.02	58.14	0.48
B12	56.94	0.70	56.60	0.45
B12	56.49	1.26	56.29	0.71
B15	93.99	2.06	92.79	1.18
B15	65.10	1.44	57.69	0.73
B17	56.92	1.43	56.63	0.74
B18	55.77	1.42	55.83	0.74
B19	1750.98	21.33	1746.28	27.28
B20	59.89	1.33	58.36	0.74
B20	66.59	1.72	57.94	0.78
B21	71.85	1.52	57.02	0.71
B22	56.40	1.33	55.49	0.71
B23	61.47	1.45	57.36	0.74
B24 ○	41.74	0.97	41.19	0.52
B24	42.51	1.06	41.88	0.54
B25	59.93	1.81	59.29	1.03
B26	56.48	2.51	56.91	1.10
B27	58.52	1.45	57.85	0.96
B27	57.76	2.13	57.14	1.05
B31	59.08	1.45	58.26	0.96
B32	57.18	1.70	55.70	0.96
B33 ○	40.71	1.16	41.01	0.69
B33	43.77	1.53	41.94	0.75
B34	55.46	1.33	55.17	0.90
B38 ○	41.23	1.72	40.69	0.76
B39	60.05	1.42	55.76	0.91
B39	55.83	1.44	56.67	0.94
B40	56.49	1.68	55.80	0.96
S1	42.36	1.09	41.00	0.99
S3	64.19	1.74	63.55	1.54
S8	7662.74	180.90	18568.34	1084.02
S13	262.94	6.27	68.25	1.69
S14	169.28	4.29	158.91	3.87
S17	57.93	1.66	56.79	1.39
S20	1898.72	22.70	1880.42	42.37
S20	1931.99	23.96	1950.37	45.60
S21	59.62	3.60	57.86	1.61
S22	82.99	2.27	61.76	1.51

S23	57.78	1.53	57.49	1.39
S24	56.06	1.41	55.82	1.34
S25	56.55	1.50	56.03	1.35
S26	57.47	1.42	56.30	1.29
S27	56.00	1.45	55.50	1.27
S28	58.70	1.89	56.55	1.34
S28	58.31	1.80	56.62	1.33
S29	76.47	2.37	58.29	1.39
S30	57.21	1.41	56.53	1.29
S31	56.07	1.50	55.76	1.29
S32	257.86	6.01	256.23	5.86
S32	254.68	5.86	252.44	5.75
S33	55.89	1.49	55.19	1.27
S33	56.58	1.74	56.37	1.33
S34	56.41	1.54	55.64	1.29
S35	67.16	2.03	56.43	1.33
S35	67.89	1.79	59.29	1.11
S36	58.01	1.63	57.28	1.08
S39	59.41	1.69	57.65	1.09
S40	58.56	1.52	57.59	1.07
S43	73.58	1.69	57.25	1.05
S44 ○	40.09	1.40	39.59	0.78
S44	42.84	1.81	39.76	0.82
S45	63.85	1.77	53.27	1.01
S46	55.07	1.46	54.73	1.02

Footnotes: 7/5 Age, $^{207}\text{Pb}/^{235}\text{U}$ age (Ma); 6/8 Age, $^{206}\text{Pb}/^{238}\text{U}$ age (Ma). Uncertainties are reported at the 2σ level. Data with circles in the left column were used calculation of the mean.

Table 5. Paleomagnetic directions obtained from the Paleo-Kuril Arc.

<i>Area</i>	Age	Location		Directions							reference	Tectonic Parameters				
Rock Unit		(°N)	(°E)	D	I	α 95	κ	n	Δ D	Δ I		R	Δ R	F	Δ F	R.P
<i>Shiranuka-hill region</i>																
Onbetsu Group	Latest Eocene to Early Oligocene	43.1	143.8	-01.4	55.1	24.0	8.8	6	45.3	24.0	Fujiwara et al. (1995)	-8.6	45.5	4.3	24.1	East Asia: 30 Ma
Urahoro Group	Late Eocene	43.1	143.8	37.3	51.5	13.9	24.3	6	22.7	13.9	this study	28.0	23.3	6.8	14.4	East Asia: 40 Ma
Urahoro Group	Late Eocene	43.1	143.8	93.7	56.8	12.3	100.9	3	22.9	12.3	Fujiwara et al. (1995)	84.4	23.5	1.5	12.9	East Asia: 40 Ma
Nemuro Group	Late Cretaceous	43.1	143.8	73.2	39.5	7.2	—	—	9.3	7.2	Hamano et al. (1986)	61.9	12.3	21.3	8.8	East Asia: 50 Ma
Nemuro Group	Late Cretaceous			66.0	84.9	22.5	3.9	15	—	22.5	Fujiwara et al. (1995)					
<i>Konsen-coastal region</i>																
Nemuro Group	Late Cretaceous	43.0	144.6	-20.8	60.0	9.4	97.5	4	14.5	9.4	Nifuku et al. (2009)	-31.2	19.8	-0.2	10.1	East Asia: 60 Ma
	Late Cretaceous to Early Eocene	43.1	145.1	0.1	61.5	13.6	15.3	9	23.3	13.6	Fujiwara and Kanamatsu (1990)	-10.4	30.0	-1.5	14.1	East Asia: 60 Ma

Footnotes: D and I, tilt-corrected declination and inclination of the observed mean directions, respectively; Δ D and Δ I, 95% confidence limit of declination and inclination, respectively. Δ D = $\sin^{-1}(\sin(\alpha 95)/\cos(I))$, Δ I = $\alpha 95$; n, number of data points used in calculating mean directions; R and F, discrepancies in declination (R) and inclination (F) between the observed and expected directions; Δ R and Δ F, 95% confidence limits of R and F, respectively. Tectonic parameters were calculated based on the definition of Beck et al. (1980); —, no data; R. P., reference pole of East Asia from Cogné et al. (2013). Paleomagnetic directions and paleomagnetic pole positions of the previous studies, except for Hamano et al. (1986), were recalculated from site-mean directions with $\alpha 95 \leq 20$. The positions of sampling sites in Fujiwara and Kanamatsu (1990) were assumed as follows: A, 43.05°N and 144.84°E; B, 43.14°N and 145.18°E; C, 43.15°N and 145.16°E; D, 43.19°N and 145.53°E; E43.15°N and 145.23°E.