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Paleomagnetism and geochronology of the Late Cretaceous-Paleogene island arc complex of the Kronotsky Peninsula, Kamchatka, Russia: Kinematic implications

Natalia M. Levashova

Geological Institute, Russian Academy of Sciences, Moscow

Mikhail N. Shapiro

United Institute of Physics of the Earth, Russian Academy of Sciences, Moscow

Vladimir N. Beniamovsky, and Mikhail L. Bazhenov

Geological Institute, Russian Academy of Sciences, Moscow

Abstract. The remnants of ancient island arcs are exposed in the Achayvayam-Valagina and East Peninsulas tectonic zones of the Olutor-Kamchatka region, northeast Russia. Geochemically, Late Cretaceous to Paleogene island arc complexes of the East Peninsulas zone are similar to intraoceanic island arcs, and the East Peninsulas zone is regarded as the extinct Kronotsky island arc. Paleomagnetic and biostratigraphic studies of Late Cretaceous and Paleogene island arc rocks were carried out at 21 sites in the eastern part of the Kronotsky Peninsula, which belongs to the East Peninsulas zone. Characteristic remanent magnetizations of both polarities isolated from most sites pass reversal and fold tests; the Cretaceous result is also supported by a positive conglomerate test on lava boulders from an intraformational conglomerate. Mean inclinations of Late Cretaceous, Ypresian and Bartonian rocks correspond to paleolatitudes of 44.8° ± 8.0°N, 38.6° ± 3.5°N, and 45.1° ± 7.0°N, respectively. These values are ~20° lower than the corresponding reference values for the North American plate for the Cretaceous and Ypresian and ~14° lower for the Bartonian. Therefore northward transport of the Kronotsky island arc is indicated. Kinematic evolution of the Kronotsky island arc in the Late Cretaceous-Paleogene was reconstructed using published kinematic parameters. In addition, we incorporated into analysis the geological and published paleomagnetic data for the Late Cretaceous and Tertiary island arc complexes of the Cape Kamchatka Peninsula from the East Peninsulas zone and the Late Cretaceous island-arc complexes of the Achayvayam-Valagina zone. As a result, two alternative scenarios of kinematic evolution for the Kronotsky island arc are proposed. According to both of them, the Kronotsky island arc was moving with the continental plate in Late Cretaceous time and with the Pacific plate from the beginning of the Paleocene until docking with the Eurasian margin in Miocene-Pliocene time.

1. Introduction

The territory between the Siberian platform and Pacific Ocean comprises numerous tectonic units of various origins and ages (Figure 1a). Currently, popular tectonic models ascribe the growth of the northwest Pacific continental margin to accretion and collision of allochthonous crustal fragments (tectonostratigraphic terranes). The Olutor-Kamchatka region in particular is thought to be composed of a number of such terranes [*Watson and Fujita*, 1985; *Zinkevich and Tsukanov*, 1992]. Most of these terranes include Late Cretaceous to Paleogene volcanic rocks of island arc affinity and are regarded as accreted remnants of island arcs. However, the number of the island arcs and their original positions and subsequent kinematics are uncertain and therefore controversial [*Kononov*, 1989; *Zonenshain et al.*, 1990; *Geist et al.*, 1994].

Combined geological and paleomagnetic studies make it possible to reconstruct the configuration of convergent boundaries for various periods of time and hence to test the alternative paleogeographic and kinematic scenarios. This seems to be a topic of major importance as the uncertainties in plate boundary locations may result in discrepancies between the observed geological data and plate motion models. Therefore new information on the tectonic evolution of northeast Asia may affect the views on plate kinematics of the entire Pacific.

The remnants of the ancient intraoceanic island arcs are found in the Achayvayam-Valagina and East Peninsulas tectonic zones of the Olutor-Kamchatka region (Figures 1b and 1c). In order to reconstruct the kinematics of these ancient island arcs one should know the interval of volcanic activity of each arc, the ages of island arc collision with the continental margin, and the paleolatitudes of each paleoarc at various times. The latter can be derived only from paleomagnetic data. Such data for the island arc complexes of the Achayvayam-Valagina zone, summarized by *Levashova et al.* [1998], indicate that these complexes had originally belonged to a single island arc which was >2000 km south from the continental margin, reaching its present position in early Tertiary time. Paleocene and Eocene paleomagnetic data for the island arc complex from Cape Kamchatka, which belongs to the East Peninsulas zone, indicate intraoceanic transport of an active island arc over the distance of >2000 km [*Pechersky et al.*, 1997]. These earlier publications, however, treated each tectonic zone separately and hence did not result in a consistent tectonic scenario for the entire Olutor-Kamchatka region.

Here we present new paleomagnetic and paleontological data for Late Cretaceous to Eocene island arc complexes from the

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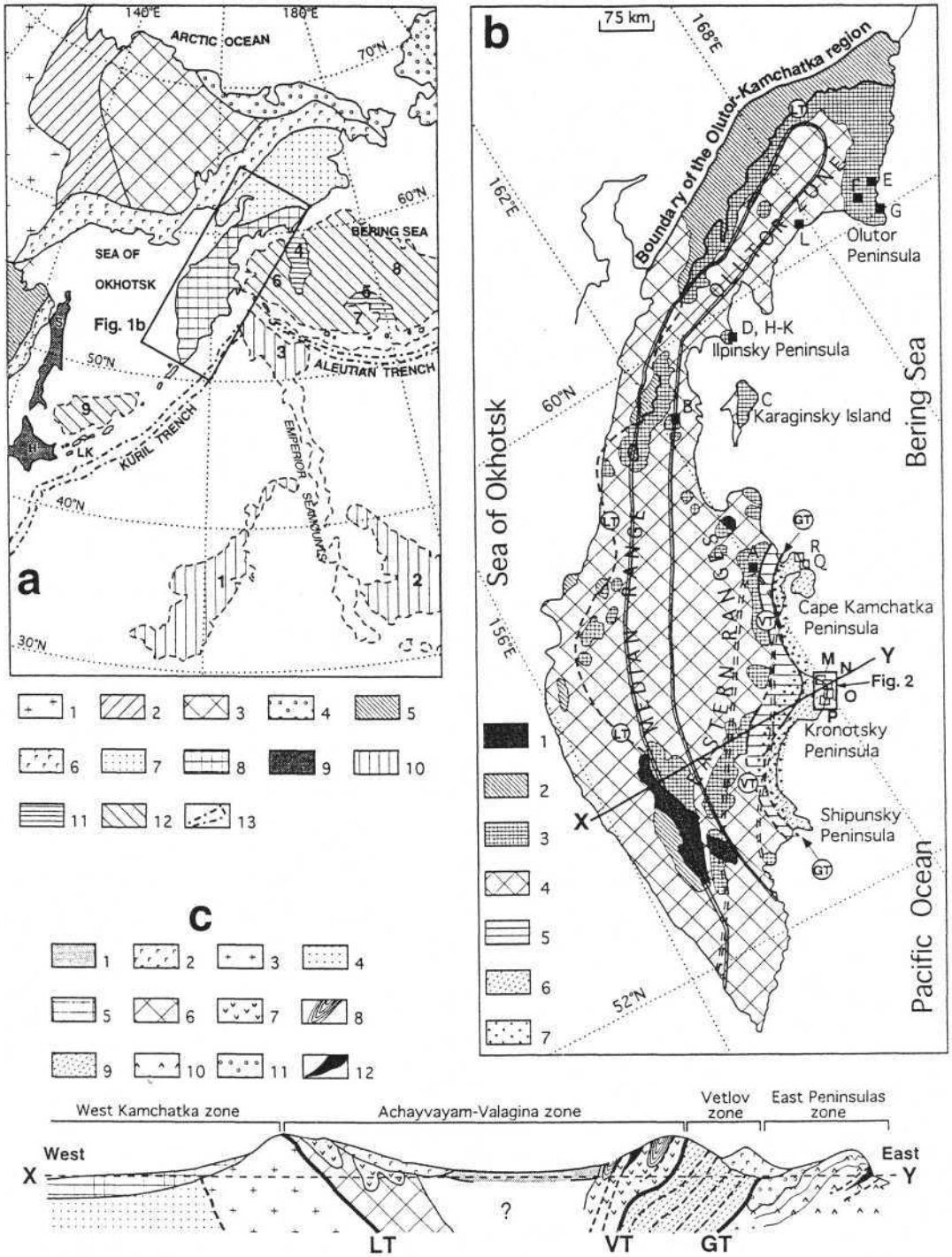

Figure 1.

Kronotsky Peninsula, which belongs to the East Peninsulas tectonic zone (Figures 1b and 2). We have also incorporated into our analysis geological data and all available paleomagnetic results from the Olutor-Kamchatka region in an attempt to reconstruct the convergent boundary kinematics between the Late Cretaceous and present.

2. Regional Tectonic Setting

The Lesnov (this fault is called the Vatyna Thrust in the Olutor zone, Lesnov Thrust on the Kamchatka Istmuth, and Andrianov Thrust in the southern part of the Median Range; for simplicity, we shall call it the Lesnov Thrust), Vetlov and Grechishkin

Thrusts are regional faults, which divide the Olutor-Kamchatka region into four main tectonic zones of the northeastern strike. These are, from the west to the east, the West Kamchatka-Ukelayat zone, the Achayvayam-Valagina zone (AVZ), the Vetlov zone and the East Peninsulas zone (EPZ) (Figures 1b and 1c).

The West Kamchatka-Ukelayat zone comprises strongly deformed sediments that accumulated on the continental slope [*Shapiro*, 1976]. These sediments are of Campanian-Maastrichtian and Late Cretaceous to early Eocene age in Kamchatka and the Olutor zone, respectively [*Chekhovich et al.*, 1990]. Late Cretaceous rocks in Kamchatka are unconformably overlain by Paleogene to Neogene coal-bearing molass and shelf sediments with minor amounts of subaerial volcanic rocks. Terrigenous formations of the West Kamchatka-Ukelayat zone are separated from Campanian-Maastrichtian volcanosedimentary rocks of the AVZ by the gentle to moderately steep Lesnov Thrust having western vergence (Figure 1c). The age of motions along this thrust becomes younger from south to north, occurring between the end of the Cretaceous to the beginning of the Paleogene in the southern part of the Median Range [*Flerov and Koloskov*, 1976] and to the Paleocene-middle Eocene on the Kamchatka Isthmus [*Shantzer et al.*, 1985]. In the Olutor zone, movement along the thrust had stopped only by middle Miocene time [*Chekhovich et al.*, 1990].

In the Achayvayam-Valagina zone to the east and southeast from the Lesnov thrust, most of the Late Cretaceous complexes are strongly deformed and cut by numerous faults. In the Olutor zone these complexes comprise Albian-Campanian mid-ocean ridge basalt and chert and Campanian-Maastrichtian tuffs and lava of island arc affinity [*Kazimirov et al.*, 1987]. The Campanian-Maastrichtian rocks are thought to be an island arc, while the Albian-Campanian series constitutes the oceanic basement of this arc [*Chekhovich et al.*, 1990]. In Kamchatka most of the Late Cretaceous volcanoclastic complexes are exposed in the Median Range, Eastern Ranges, and on Karaginsky Island (Figure 1b). The Median Range and Eastern Ranges are separated by the sediment-filled late Cenozoic Central Kamchatka basin. The island arc complexes are unconformably overlain by Eocene to Oligocene marine sediments and Miocene to Quaternary volcanic rocks. In the Eastern Ranges, however, the island arc rocks are overlain conformably and with a gradual transition by middle-upper Paleocene terrigenous flysch. A continent-derived material is absent altogether in volcanic rocks; in contrast, volcanic material of island arc affinity is very scarce in the flysch [*Bakhteev et al.*, 1994]. Geochemically, the Late Cretaceous volcanoclastic complexes of the entire AVZ are similar to volcanic rocks of intraoceanic island arcs such as the Tonga arc [*Zinkevich et al.*, 1993]. These complexes were assumed to originally belong to the single Achayvayam-Valagina island arc (AVIA) [*Zinkevich and Tzukanov*, 1992]. This hypothesis was confirmed by paleomagnetic data on the island arc complexes of the entire AVZ [*Levashova et al.*, 1998].

The steep Vetlov Thrust separates the AVZ and Vetlov zone, which consists of Paleocene to Eocene siliceous and terrigenous sediments with some lenses of MORB-type basalts, pelagic cherts and limestones; at least a part of these lenses belongs to chaotic complexes [*Tsukanov*, 1991; *Bakhteev et al.*, 1994]. Some cherts contain Late Cretaceous radiolaria. The Vetlov zone is strongly deformed and cut by numerous steep thrusts of eastern vergence. This zone is regarded as an accretionary prism [*Zinkevich and Tsukanov*, 1992] which was formed after the AVIA had docked with the continental margin and the Oligocene-Miocene subduction-related volcanic activity had commenced in central Kamchatka (Figure 1b).

The East Peninsulas zone mainly comprises Paleocene-Eocene volcanics and volcanoclastics exposed at Cape Kamchatka and Kronotsky and Shipunsky Peninsulas (Figure 1b), whereas Late

Figure 1. (opposite) (a) Main tectonic units of the northwest Pacific and adjacent areas. 1, Siberian platform; 2, Verkhoyansk fold belt; 3, Kolyma-Omolon folded domain; 4, Chukotka-North Alaska fold belt; 5, Mongol-Okhotsk and Sikhote-Alin fold belts; 6, Okhotsk-Chukotka (Cretaceous) and Sikhote-Alin (Late Cretaceous-Paleocene) volcanic belts of the Andean type; 7, North Koryak amalgamated terranes; 8, Olutor-Kamchatka fold belt; 9, Hokkaido-Sakhalin fold belt: S, H, and LK, Sakhalin, Hokkaido, and the Lesser Kuril Islands; 10, Oceanic uplifts (1, Shatsky; 2, Hess; 3, Obruchev); 11, submarine ridges of the Bering Sea (4, Shirshov; 5, Bauers); 12, deep basins of the marginal seas (6, Komandorsky; 7, Bauers; 8, Aleutian; 9, South Kuril); 13, trenches. (b). Schematic geological map of Kamchatka. 1, pre-Cretaceous metamorphic rocks; 2, Late Cretaceous continent-derived terrigenous rocks of the West Kamchatka-Ukelayat zone; 3, Late Cretaceous island arc complex of the Achayvayam-Valagina zone (AVZ); 4, Cenozoic rocks of the West Kamchatka-Ukelayat zone and AVZ; 5, Vetlov zone; 6, Late Cretaceous-Eocene island arc complex of the East Peninsulas zone (EPZ); 7, Tushev basin. Thick lines are main thrusts: LT, Lesnov Thrust; VT, Vetlov Thrust; GT, Grechishkin Thrust. Solid and dashed segments denote mapped and inferred parts, respectively. Double solid and dashed lines are the boundaries of the Oligocene-Miocene subduction-related volcanic belt of central Kamchatka and the Pliocene-Quaternary subduction-related volcanic belt of east Kamchatka, respectively. Solid squares are sampling localities in the AVZ: A, *Levashova et al.* [1997]; B, *Levashova et al.* [1998]; C-L, *Kovalenko* [1992, 1993] and *Kovalenko et al.* [2000]. Open squares are sampling localities in the EPZ: M-O, this study; P-Q, *Pechersky et al.* [1997]. (c) Schematic cross section of Kamchatka along X-Y line (Figure 1b). 1, Pliocene-Quaternary sediments in the Central Kamchatka basin; 2, Neogene-Quaternary subaerial volcanics of the central Kamchatka and east Kamchatka volcanic belts; 3-5, West Kamchatka zone: 3, pre-Cretaceous metamorphic rocks; 4, Late Cretaceous flysch; 5, Eocene to Pliocene molasses; 6-8, Achayvayam-Valagina zone: 6, Santonian-Campanian basalt and chert; 7, Campanian to lower Paleocene volcanic rocks; 8, upper Paleocene terrigenous flysch; 9, upper Paleocene to Miocene Vetlov accretionary complex; 10, Late Cretaceous to Eocene island arc complex of the East Peninsulas zone; 11, Oligocene to Miocene shelf sediments; 12, ophiolites. Thick solid lines are main faults: LT, Lesnov Thrust; VT, Vetlov Thrust; GT, Grechishkin Thrust. Thick dashed lines are other faults.

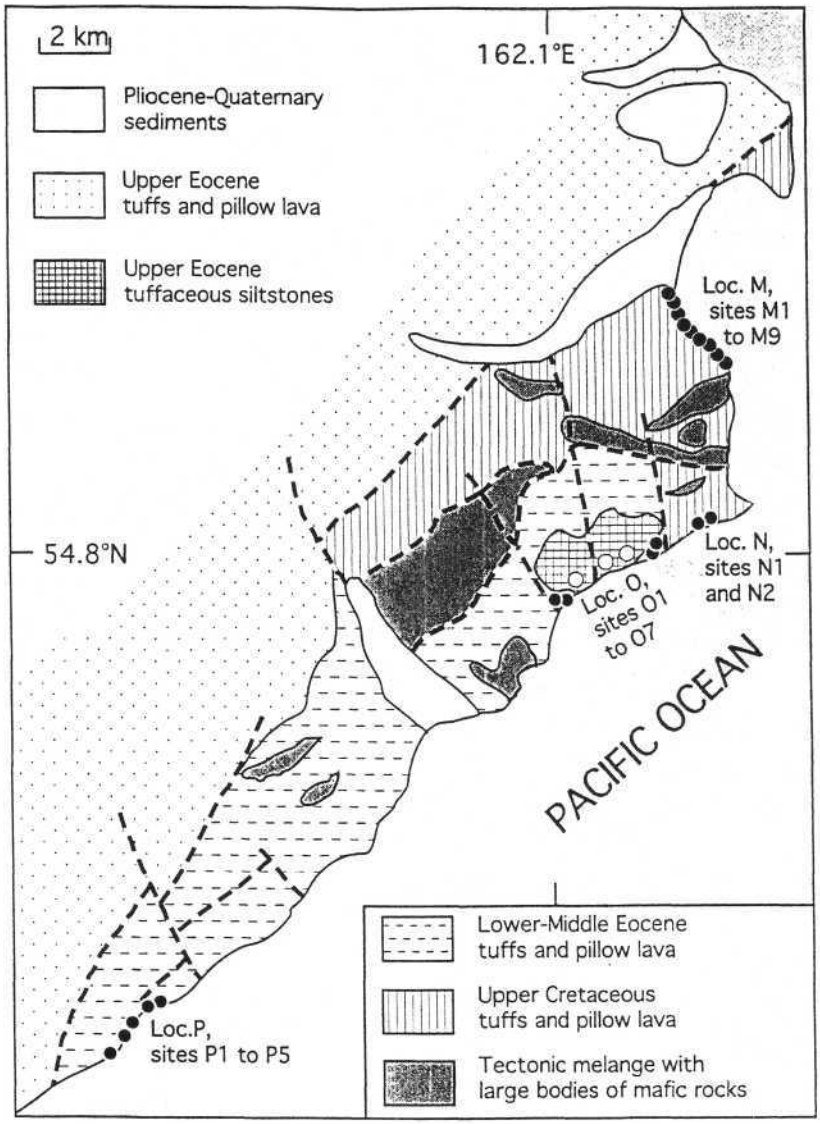

Figure 2. Schematic geological map of the northeastern part of the Kronotsky Peninsula and sampling sites of this study (solid dots) and those from *Bazhenov et al.* [1992] (open dots).

Cretaceous basalts and tuffs of island arc affinity are found in a limited part of the Kronotsky Peninsula (Figure 2). Paleogene basalts are widespread on the Kronotsky Peninsula and very similar to those from Cape Kamchatka [*Khubunaya*, 1987]. All basalts are of island arc affinity; MORB-type and intraplate basalts are unknown here. Irrespective of their age, high-alumnium plagiotholeiites from the EPZ resemble plagioclase-porphyric tholeiites from the Tonga, Kermadec, and other island intraoceanic arcs [*Khubunaya*, 1987; *Puzankov*, 1994]. Paleogene rocks of the EPZ are significantly different from contemporaneous rocks of the other tectonic units of Kamchatka, where lower Eocene rocks are mainly siliceous and middle to upper Eocene rocks are terrigenous. On the whole, the EPZ is regarded as the extinct Kronotsky island arc [*Zinkevich and Tsukanov*, 1992]. Although Late Cretaceous and Paleogene island arc complex is deformed everywhere in the EPZ, the age of deformation is poorly constrained owing to a lack of overlying well-dated rocks. In particular, folding on the Kronotsky Peninsula is older than Pliocene.

The EPZ is separated from the Kamchatka mainland by the Tushev basin filled with upper Eocene to Miocene terrigenous sediments that are best exposed to the south of the Kronotsky Peninsula. The steep Grechishkin Thrust of eastern vergence (Figure 1c) divides the basin into two parts with different section types: the western part includes the youngest member of the Vetlov accretionary complex, whereas sediments of the eastern part overly Eocene volcanics of the Kronotsky Peninsula with erosional and weak angular unconformity. The Grechishkin Thrust is regarded as a suture which resulted from the late Miocene collision of the Kronotsky block with Kamchatka [*Bakhteev et al.*, 1997]. This thrust and conjugate folds in the Tushev basin are overlain with major angular unconformity by flat-lying Pliocene volcanics. This thrust can be traced to the Cape Kamchatka Penin-

sula and to an area west of the Shipunsky Peninsula; almost everywhere this structure, however, is hidden under either Pliocene-Quaternary volcanics, sediments, or water.

Middle Eocene deformation affected almost all of Kamchatka, with only minor deformation occurring in the Miocene and Pliocene. In contrast, the major folding and thrusting are of Miocene age in the Tushev basin. Paleocene-Eocene island arc volcanics are abundant in the EPZ, but very scarce inland. Widespread Oligocene-Miocene subduction-related volcanics in central Kamchatka partly overlap the West Kamchatka-Ukelayat zone and AVZ, and similar Pliocene-Quaternary volcanism is mainly limited to east Kamchatka (Figure 1b).

3. Geological Setting and Sampling

Paleogene volcanic rocks on the Kronotsky Peninsula are weakly deformed, constituting a regional westward dipping homocline cut by numerous faults with small conjugate folds. Late Cretaceous rocks are limited to the northeastern part of the Peninsula and are more deformed; for instance, these strata form an anticline with the partly overturned northern limb at locality M (Figure 2). Everywhere, Late Cretaceous and Paleogene rocks are separated by melange zones and/or faults with intense conjugated folds. On the other hand, Late Cretaceous and Paleogene formations are visibly very similar, and their subdivision is based on scarce fauna, thus making unclear the exact relationship between Late Cretaceous and Paleogene rocks.

The oldest part of the Late Cretaceous section (lower member of the Cape Kamenisty Formation [*Raznitsyn et al.*, 1985]) is composed of basalt flows up to several tens of meters thick intercalated with layered coarse- to fine-grained tuffs and conglomerates with abundant lava boulders indicating shallow water accumulation for these rocks. Fine-grained varieties are finely stratified, without traces of slumping. All contacts of this member with other formations are faulted. These rocks form an anticline and are locally intruded by small diabase bodies of unknown age. At the limbs of this fold (locality M; Figure 2), 105 oriented hand samples were collected from volcanoclastic rocks (seven sites) and pillow lava (five sites). In addition, 18 lava boulders from an intraformational conglomerate were collected for the conglomerate test. According to radiolaria found at the section base (*Orbiculiforma quadrata, Orbiculiforma cf. vacaensis, Archaeospongoprunum* sp. (sp 1, sp 2, sp 3), *Prunobrachium ex gr. sibericum, Lithocampe aff. eligantissima, Dictyocephalus* sp., and *Archaedictyomitra* sp.), these rocks are of Coniacian-early Campanian age [*Tsukanov*, 1991]. Later, new radiolaria found in the tuffaceous rocks of the same section (*Porodiscus cretaceous Clark et Campbell, Amphibrachium sibericum Gorbovetst, Prunobrachium sibericun (Lipman), P.cf. crassum (Lipman), Spongurus* sp., *Stylosphaerrs cf. pussilla (Campbell et Clark), Staurodictya fresnoensis Foreman, Prtoxiphotractus* sp., *Archaeodictyomitra regina (Campbell et Clark), Dictyomitra striata Lipman, D. multicostata Zittel gr.*, and *Amphipyndax cf. alamedaensis (Campbell et Clark)* (M. Boyarinova, unpublished data, 1997) and *Orbiculiforma renillaeformis Pessagno, Amphibrachium cf. mucronatum Lipmam, Spongrurus sp., Prunopyle* sp., *Prunobrachium incisum Koslova, P. cf. longum Pessagno, Dictyomitra lamellicostata Foreman, Strictomitra manifesta Foreman*, and *Amphipyndax stocki (Campbell et Clark)* (V. Vishnevskaya, unpublished data, 1996)) indicate a Campanian-Maastrichtian age for the section.

The upper member of the Cape Kamenisty Formation [*Raznitsyn et al.*, 1985], exposed at locality N (Figure 2), is cut by numerous faults. Twenty-five samples were collected from the 200 m thick section of pillow lava (site N1) and tuffs (site N2). Poorly preserved radiolaria (*Orbiculiforma renillaeforumis, Amphibrachium cf. mucronatum, Prunobrachium? incisum, Spongurus* sp., and *Prunopyle* sp.) indicate a Maastrichtian-Paleocene age for the tuffaceous rocks [*Tsukanov*, 1991]. However, an Ar/Ar determination on a single sample of pillow lava (*P. Renne*, unpublished data, 1996) indicates an age of ~80 Ma (early Campanian). On the whole, the age of the rocks studied at localities M and N is most likely Campanian-Maastrichtian.

Eocene volcanoclastic rocks underlain by basal conglomerates and pillow lava form a gentle syncline at locality O (Figure 2). The upper part of this section was studied by *Bazhenov et al.* [1992]; for this study, twenty-nine oriented hand samples were collected from pillow lava at four sites. Planktonic foraminifera (*Subbotina aequiensis, S. aff. contorta, S. cf. nana, S. aff. incisa*, and *S. ex gr. achtschacujmensis*) at the base of the sedimentary section at locality O (Figure 2) can be correlated with the early Ypresian (early Eocene) zone *Morozovella wilcoxensi* which has been recognized in other sections of the West Pacific [*Beniamovsky et al.*, 1992; *Volobueva et al.*, 1994]. A few meters above, another complex of planktonic foraminifera (*Subbotina inaequispira, S. pseudoeocaena pseudoeocaena, S. cf. eocaenica*, and *Globanimalina (Pseudohastigerina) wilcoxensis*) is similar to the *Globanomalina (Pseudohastegerina) wilcoxensis* zone in upper lower Ypresian to lower Lutetian (middle Eocene) rocks of the western North Pacific [*Beniamovsky et al.*, 1992; *Beniamovsky and Gladenkov*, 1996]. Still a few meters higher, a rich complex of foraminifera (*Subbotina inaequispira, S. pseudoeocaena pseudoeocaena, Pseudohastigerina wilcoxensis, Subbotina boweri, S. linaperta, S. peudoeocaena compacta, S. posttriloculinoides, S. eocaenica irregularis, Pseudohastigerina micra, Acarinina cf. bullbrooki*, and *A. broedermanni*) belongs to the *Subbotina boweri* zone and indicates a middle Lutetian age for this horizon. The three zones mentioned above comprise the lower 10-13 m of the section. According to the foraminifera (*Subbotina praebulloides, S. incretacea, S. galavasi ,S. tripartita, Catapsidrax dissimilis*, and *Acarinina rugosoaculeata*), the upper part of the section belongs to the *Subbotina praebulloides* zone of Bartonian (late Eocene) age that is well known in east Kamchatka [*Beniamovsky et al.*, 1992; *Beniamovsky and Gladenkov*, 1996].

Thus the sediments at locality O cover the interval from the early Ypresian to Bartonian, which is in agreement with the available nannoplankton data [*Shcherbinina*, 1997]. The first three zones, however, comprise <15 m of the section, whereas the Bartonian part is several tens of meters thick. Therefore the main part of the collection of *Bazhenov et al.* [1992], who mainly sampled the middle and upper parts of the sedimentary sequence at this locality, is of Bartonian age (ca. 40 ± 2 Ma), whereas the section base and underlying basalts are older. Thus the basalts are probably not younger than 56-53 Ma. In sharp contrast to these data is the Ar/Ar determination on a single basalt sample of ~40 Ma (*P. Renne*, unpublished data, 1996).

Gently southwest dipping Eocene intercalated pillow lava and volcanoclastics were studied at locality P (Figure 2). Thirty-nine samples of pillow lava and volcanoclastics were taken at five sites. Benthic foraminifera of the *Subbotina praebulloides* zone indicate a Bartonian age for the sediments. Therefore we conclude

Table 1. Paleomagnetic Data From the Late Cretaceous Rocks (Localities M and N)

				In Situ				Tilt corrected				
Site	N/N_0	T	B	D	I	k	a_{95}	D	I	k	a_{95}	P
M1	5/5	5	114/30	88.4	-27.6	17	15.4	74.2	-53.5	17	15.1	R
M2	5/5	6	114/30	104.4	-39.8	14	16.9	92.6	-69.3	14	16.7	R
M3	6/7	5	106/27	102.7	-22.3	10	18.3	99.2	-59.3	10	17.7	R
M4	12/18	35	122/28	130.2	-40.6	9	13.8	139.1	-68.4	10	13.0	NR
M5	5/5	15	131/13	137.1	-54.3	27	12.1	140.1	-66.9	26	12.9	R
M6	9/11	10	111/18	113.5	-45.6	6	19.9	116.0	-62.8	7	17.3	NR
M7	7/9	10	307/19	78.8	-79.0	4	25.1	112.6	-62.8	6	21.9	NR
M8	8/10	12	300/72	328.9	-45.3	16	12.5	85.0	-53.1	17	11.9	NR
M9	10/12	40	314/87	320.9	-40.9	5	19.4	120.9	-52.9	13	12.2	NR
M10[a]	6/6	15	15/11	224.5	62.8	50	8.1	241.5	71.8	47	8.3	N
M11[a]	6/6	15	17/10	219.7	56.4	47	8.4	227.7	65.4	46	8.5	N
M12[a]	8/11	12	45/8	234.4	67.6	30	9.1	240.3	75.9	40	7.8	N
N1	5/11	20	115/8	260.6	54.5	18	15.1	253.1	60.7	21	13.8	N
N2	6/14	15	157/22	306.6	50.8	8	20.2	283.7	67.7	27	11.1	N
Mean	78/107			97.4	-61.8	3	84.4	106.8	-63.1	10	4.9	
	$F^b_{(4,150)} = 2.42$				$f = 34.81$					$f = 1.84$		
Mean[c]	(11/14)			97.6	-59.5	5	19.1	103.1	-63.3	43	6.4	
	$F^d_{(4,14)} = 3.00$				$f = 9.98$					$f = 0.84$		

Sites are labeled as in text. N, the number of the samples (sites) studied/accepted; T, true thickness studied (in meters); B, dip azimuth/dip angle (in degrees); D, declination (in degrees); I, inclination (in degrees); k, concentration parameter; a_{95}, radius of confidence circle (in degrees); P, polarity of the ChRM: N, normal, R, reversed, and NR, mixed; F, the 95% critical value of F statistic; f, calculated value of the same statistic. Superscripts a and b are for the fold test at the sample and site levels, respectively. Underlined result was used for interpretation.

[a] These data were not used for computation of overall mean direction.

[b] Data are for the fold test at the sample level.

[c] These results in stratigraphic coordinates were used for interpretation.

[d] Data are for the fold test at the site level.

that the rocks studied at locality P and the sedimentary rocks from locality O, studied by *Bazhenov et al.* [1992], are of Bartonian age (ca. 40 ± 2 Ma).

4. Paleomagnetic Study

4.1. Methods

The thicknesses of the outcrops (sites) studied vary from several meters to several tens of meters (Tables 1 and 2), and samples were distributed more or less uniformly across the outcrop. One hand sample oriented with a magnetic compass was taken at each stratigraphic level. One to two cubic specimens from each hand sample were subjected to progressive thermal demagnetization in 12-18 steps up to 610°C. One specimen was thermally demagnetized in a homemade oven with internal residual fields of ~10 nT and measured with a JR-4 spinner magnetometer having a noise level of 0.05 mA/m in the Paleomagnetic Laboratory of the Geological Institute in Moscow. Some duplicate specimens were thermally demagnetized in a multishield furnace and measured with a cryogenic magnetometer in a nonmagnetic room in the Paleomagnetic Laboratory of the Institut de Physique du Globe in Paris. Duplicate specimens from some basalt samples were subjected to alternating field demagnetization with a 2G cryogenic magnetometer in the Paris Laboratory. Anisotropy of magnetic susceptibility (AMS) was measured with the aid of a kappa-bridge KLY-2. Demagnetization results were plotted on orthogonal vector diagrams [*Zijderveld*, 1967], and linear trajectories were used to determine directions of magnetic components by a least squares fit comprising three measurements or more [*Kirschvink*, 1980]. The characteristic remanent magnetization (ChRM) was determined without anchoring the final linear segments to the origin of the vector diagrams. Components isolated from the sister specimens of a hand sample were used to calculate the sample means, which were in turn used to calculate site means.

4.2. Results

4.2.1. Late Cretaceous rocks. Pillow basalt and tuffaceous rocks at locality M are strongly magnetized, with intensities of natural remanent magnetization (NRM) ranging from 1.63 to 71.9 A/m in the basalts and from 0.11 to 11.4 A/m in the sediments. A single ChRM was isolated from most lava samples above 200° (Figure 3a); in just a few samples, a low-temperature overprint persisted up to 400° (Figure 3b). Some tuff samples also had a single ChRM above 200°-300° (Figure 3c), whereas another component is present at low to intermediate temperatures in many samples (Figures 3d and 3e). This intermediate-temperature component has a direction statistically indistinguishable from the ChRM direction in stratigraphic coordinates but is much more scattered. Although represented by linear segments on demagnetization plots, this component is likely to result from partial overlap of unblocking spectra of a recent overprint and the ChRM. Some samples, after a sharp initial drop in NRM intensity, showed erratic behavior above 350°-400°C and were rejected. Unblocking temperatures for most samples between 550° and 580° indicate that magnetite is the main remanence carrier (Figure 3). Magnetizations of both polarities were found, and an upsection R-N-R-N succession of polarity zones was identified (R indicates reversed, and N indicates normal).

The NRM intensity ranges from 1.57 to 16.2A/m in basalt and from 0.09 to 0.93 A/m in sediments at locality N (Figure 2). Many samples from this locality yielded no consistent data after removal

Table 2. Paleomagnetic Data From the Eocene Rocks (Localities O and P)

Site	N/N_0	T	B	In Situ D	In Situ I	In Situ k	In Situ a_{95}	Tilt corrected D	Tilt corrected I	Tilt corrected k	Tilt corrected a_{95}
					Ypresian Data						
O4	7/8	20	216/23	316.8	60.1	21	11.5	278.3	56.3	21	11.6
O5	8/8	35	215/23	322.9	65.4	11	15.0	274.7	62.5	11	15.2
O6	5/7	20	40/20	261.1	45.7	36	10.5	281.4	58.1	35	10.7
O7	4/6	15	40/21	261.8	41.5	70	8.4	279.7	54.7	71	8.3
Mean	24/29			290.8	59.0	10	9.0	278.2	58.4	20	6.4
	$F^a_{(2,44)} = 3.21$				$f = 22.74$				$f = 0.22$		
Mean[b]	(4/4)			283.6	56.5	13	19.6	278.7	57.9	49	3.2
	$F^c_{(2,24)} = 6.94$				$f = 81.23$				$f = 0.97$		
					Bartonian Data						
O1	13/14		17/8	286	61	18	9.2	301	60	18	9.2
O2	17/22		222/21	332	61	16	8.6	292	61	19	7.8
O3	9/10		190/10	324	54	32	8.2	312	60	29	8.6
P1	8/9	30	294/14	323.5	75.7	27	9.5	309.3	62.7	28	9.5
P2	7/8	30	295/14	210.9	87.0	16	13.2	283.0	75.1	17	12.9
P3	11/11	35	296/16	229.5	76.7	11	12.9	265.8	65.4	11	12.8
P4	7/7	15	306/17	307.2	81.8	54	7.2	306.7	65.2	60	6.8
P5	4/4	10	285/44	66.8	80.1	28	13.3	294.7	53.5	40	11.1
Mean	76/85			310.4	72.9	12	4.6	296.0	63.5	18	3.8
	$F^a_{(10,140)} = 1.90$				$f = 8.96$				$f = 1.46$		
Mean[b]	(8/8)			311.6	76.5	22	10.5	296.8	63.5	78	5.6
	$F^c_{(2,12)} = 3.89$				$f = 9.58$				$f = 0.79$		

Sites O1 to O3 are from *Bazhenov et al.* [1992]. Other notation is as in Table 1.
[a] Data are for the fold test at the sample level.
[b] These results in stratigraphic coordinates were used for interpretation.
[c] Data are for the fold test at the site level.

of a low-temperature component between 250° and 300°C, and less than a half of tuff (Figure 3h) and basalt (Figure 3i) samples retained a ChRM at higher temperatures. This ChRM is unblocked at 540°-570°C, indicating that magnetite is the principal carrier of remanence. All ChRM directions are of normal polarity at this locality (Table 1).

The site mean directions from localities M and N are much better clustered after tilt correction than in situ, and tilt-corrected ChRM of both polarities are nearly antipodal (Figure 4); nevertheless, the data do not pass the fold test [*McFadden and Jones*, 1981]. We suspect that the data set may be somehow distorted: for instance, it may be contaminated by secondary components. A closer inspection shows that the ChRM directions at sites M10-M12 are grouped much better than those of the main collection (concentration parameter k = 41.6 and 10.3, respectively; Table 1). The mean direction for the samples from these three sites of declination D = 226.3°, inclination I = 63.2°, and radius of confidence circle a_{95} = 5.0° in situ and D = 235.7°, I = 71.6°, and a_{95}= 4.9° after tilt correction differs in both declination and inclination from the other data; the fold test for these three sites is inconclusive. Along the coastline of Kronotsky Peninsula, diabase intrusions of unknown age are exposed, one of them cropping out close to sites M10-M12. We infer that these sites were remagnetized by the intrusion and prefer to exclude them from further analysis. For the remaining data the fold test [*McFadden and Jones*, 1981] is positive at the 95% confidence level on both the site mean and sample levels for localities M and N combined (Table 1).

The NRM in the basalt boulders is mainly accounted for by a single component showing linear decay to the origin, similar to samples of the host basalt (Figure 3h); in just a few samples, another component was removed below 350°C (Figure 3i). The ChRM directions are randomly distributed (Figure 4c), yielding the normalized vector resultant of 0.159, which is much less than the 95% critical limit of 0.377 [*Mardia*, 1972].

Thus the ChRM in Late Cretaceous rocks passes the fold test, and its normal and reversed directions are nearly antipodal. The conglomerate test is also positive for the rocks from locality M. Relatively thick stratigraphic intervals studied at the accepted sites, the tight grouping of tilt-corrected site means (Figure 4b, and Table 1), and the presence of several polarity zones testify to adequate averaging of geomagnetic secular variation. Close agreement of ChRM directions in the lava flows and sediments indicates that no inclination error is present. At some sites we sampled well-stratified fine-grained tuffaceous rocks which most likely would not accumulate on a tilted surface. Therefore the ChRM in the Late Cretaceous rocks is most probably primary and a reliable record of the ancient geomagnetic field.

4.2.2. Eocene rocks. Ypresian basalts at locality O (Figure 2) are strongly magnetized, and NRM intensities range from 3.93 to 70.4 A/m. After removal of a weak scattered overprint by 250°-300°C, a stable ChRM of ubiquitously normal polarity, represented by linear trajectories decaying to the origin above 300°-350°, was isolated from most samples (Figure 5a). Alternating field demagnetization gave the same results (Figure 5b). The ChRM persisted up to 570°-580°, implying that magnetite is the ChRM carrier. The ChRM directions are well grouped within each site, and the site means are well defined (Figures 6a and 6b and Table 2). The fold test [*McFadden and Jones*, 1981] is positive at the 95% confidence level (Table 2) indicating a prefolding origin of the ChRM. AMS in these basalts is always <3%, and the axes

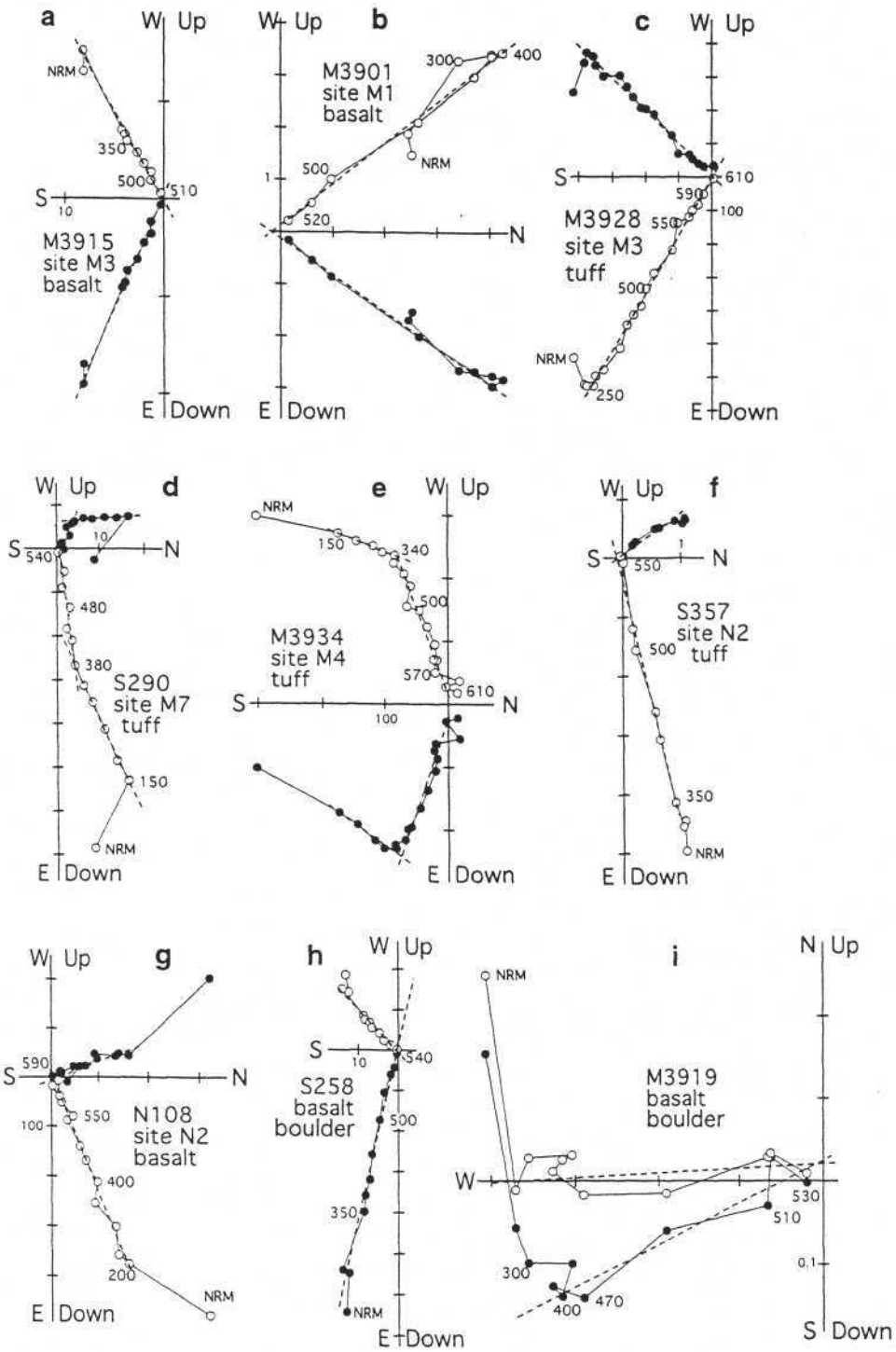

Figure 3. Representative vector component plots for Late Cretaceous basalt flows (Figures 3a, 3b, and 3g) and tuffaceous rocks (Figures 3c-3f) from localities M (Figures 3a-3e) and N (Figures 3f and 3g) and basalt boulders from intraformational conglomerate (Figures 3h and 3i). Solid (open) circles are projected onto the horizontal (vertical) plane. The plots are in stratigraphic coordinates. Isolated components are shown as dashed lines. Temperatures are in degrees Celsius. Magnetizations are in A/m for basalt samples (Figures 3a, 3b, and 3g-3i) and mA/m for tuffs (Figures 3c-3f).

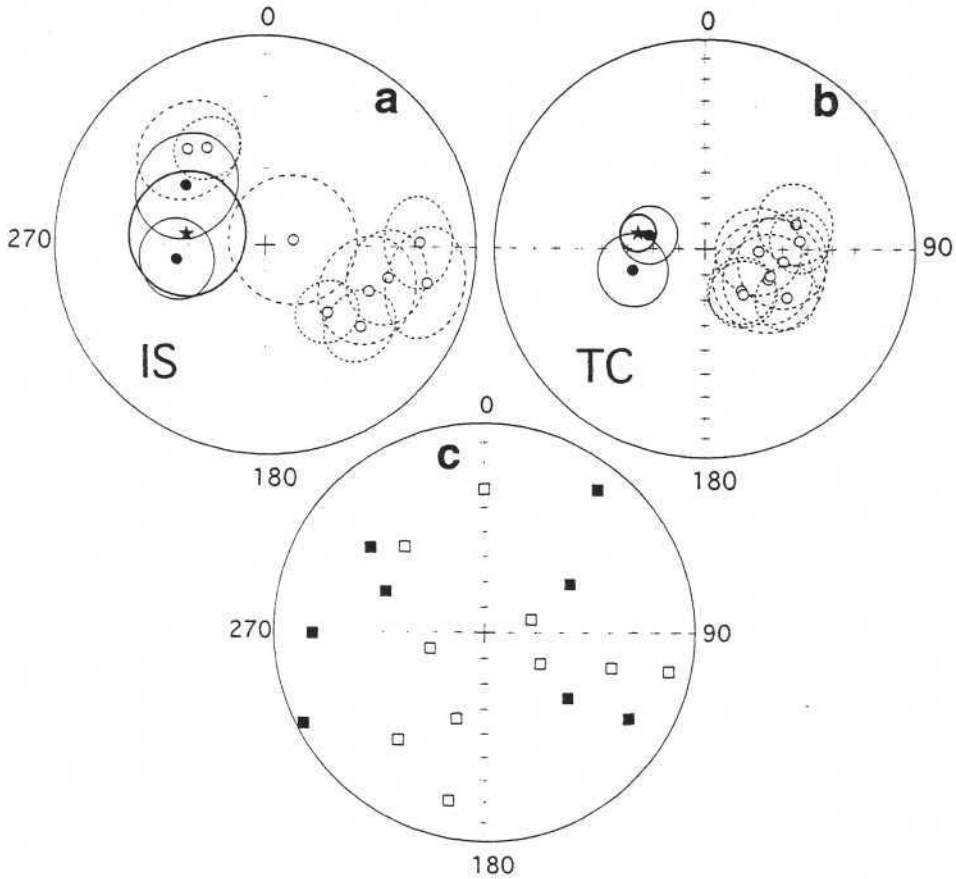

Figure 4. Stereoplots of site mean directions (dots) and overall mean direction (star) with associated confidence circles from Late Cretaceous rocks from localities M and N (a) in situ and (b) after tilt correction. (c) Stereoplot of ChRM directions (squares) for lava boulders from Late Cretaceous intraformational conglomerate. Solid (open) symbols and solid (dashed) lines are projected onto lower (upper) hemisphere.

of the AMS ellipsoid are randomly distributed on the unit sphere. This implies that the ChRM directions are not distorted despite very high NRM intensities in basalts.

NRM intensities range from 1.89 to 30.7 A/m in basalts and from 150 to 420 mA/m in Bartonian sediments at locality P (Figure 2). After removal of a low-temperature overprint below 300°, a ChRM was isolated from most samples (Figures 5c and 5d). The ChRM persisted to 550°-560°C, and sometimes to 600°C, indicating that magnetite is the carrier. The ChRM is of normal polarity in all samples but two; the latter were inverted through the origin and included into our analysis.

Bazhenov et al. [1992] reported positive fold and reversal tests for paleomagnetic data from Bartonian sedimentary rocks just above the basalts from locality O. For combined data from Barto-

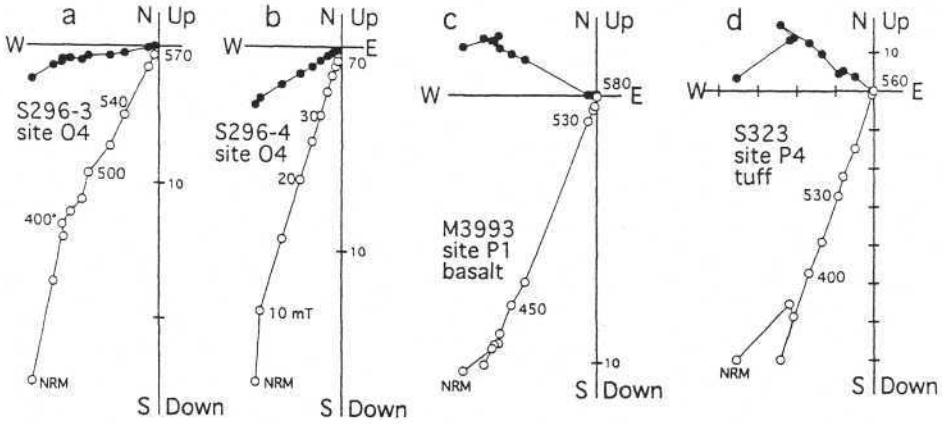

Figure 5. Representative vector component plots for (a-b) Eocene lava from locality O and (c-d) volcanics from locality P. For Figure 5b, steps are in millitesla. Other notation is as in Figure 3.

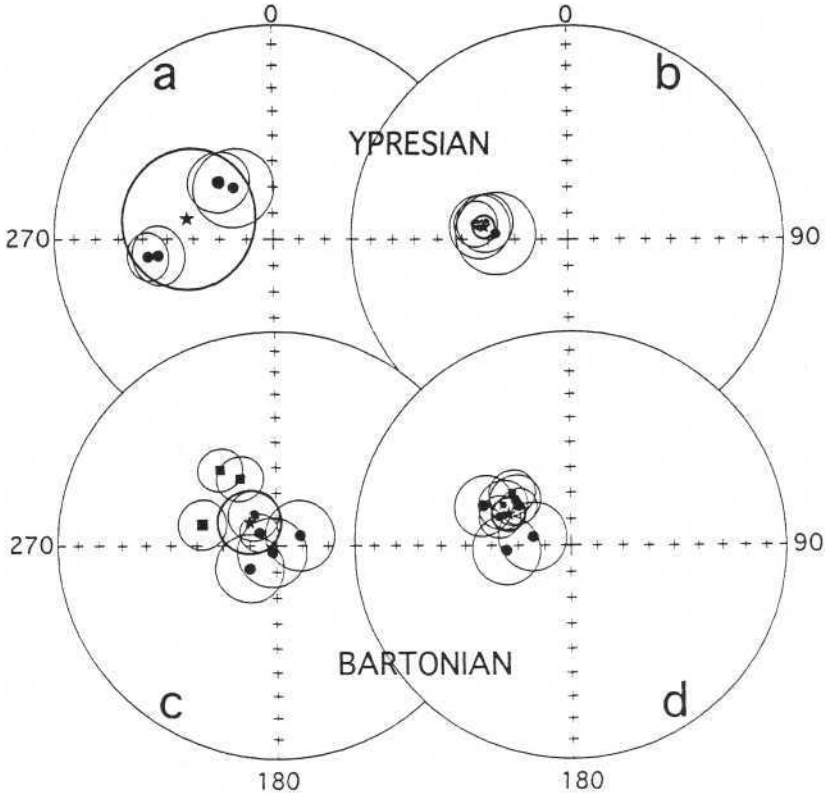

Figure 6. Stereoplots of site mean directions (dots, this study; squares, *Bazhenov et al.* [1992]) and overall mean directions (stars) with associated confidence circles from (a,b) Ypresian and (c,d) Bartonian rocks from localities O and P in situ (Figures 6a and 6c) and after tilt correction (Figures 6b and 6d). Same notation appears as in Figure 4.

nian rocks from localities O and P the site means are well grouped after tilt correction (Figures 6c and 6d), and the fold test [*McFadden and Jones*, 1981] is positive on both the sample and site mean levels at the 95% confidence level (Table 2). Relatively thick stratigraphic intervals studied at each site and the presence of polarity zones mean that secular variation is adequately averaged. Therefore the ChRM is most probably of primary origin in the Bartonian rocks, and its mean direction is reliably established.

5. Summary of Paleomagnetic and Rock Ages Data

In most samples, radiolaria indicate a Campanian-Maastrichtian age for the rocks at locality M. However, some of the samples were dated as Coniacian-early Campanian. Paleomagnetic study revealed an R-N-R-N polarity sequence (from bottom to top) at locality M; thus at least two zones of normal and two of reversed polarity are present here. The upper boundary of the Cretaceous normal superchron at ca. 83 Ma coincides with the Santonian-Campanian boundary [*Cande and Kent*, 1992], and the pre-Campanian age of the studied section can be excluded. Moreover, there are only one reversed and one normal polarity chrons in Campanian time [*Cande and Kent*, 1992], and it is unlikely that the 200 m thick studied section of intercalated lava flows, conglomerates, and tuffs encompasses the two oldest long R and N polarity zones in the Campanian (ca. 8 Ma). We think that the observed R-N-R-N polarity sequence is probably not older than early Maastrichtian (anomaly 32, 72-74 Ma) when shorter polarity zones occurred. Therefore we infer the age of the section to be late Campanian-Maastrichtian or even Maastrichtian.

Poorly preserved radiolaria imply a Maastrichtian-Paleocene age for the rocks at locality N. An Ar/Ar age for basalt from the lower part of the section is ~80 Ma. However, all ChRM directions from this locality are of normal polarity, while the field was reversed between 83 and 79 Ma [*Cande and Kent*, 1992]. Therefore this section accumulated either before 83 Ma or between 79 and 74 Ma. Judging from the paleontological data, the later interval is more probable. Tentatively, we assume that the rocks from localities M and N accumulated in the late Campanian-Maastrichtian; for this reason we pooled the paleomagnetic data for localities M and N. The Late Cretaceous overall mean direction of D = 283.1°, I = 63.3°, and a_{95} = 6.4°, corresponds to a paleolatitude of 44.8° ± 8.0°N; we assign a 73 ± 7 Ma age to this result.

Paleontological data indicate a Bartonian age (42-38 Ma) for sediments from locality O and all rocks from locality P. The overall mean direction of D = 296.0°, I = 63.5°, and a_{95} = 5.6°, for these rocks corresponds to a paleolatitude of 45.1° ± 7.0°N. The presence of both polarities in Bartonian rocks at locality O [*Bazhenov et al.*, 1992] implies that a considerable time interval was spanned; we cannot, however, correlate these zones with a paleomagnetic timescale and have to assign the age of 40 ± 2 Ma to this result.

A tilt-corrected mean direction for basalts at locality O of D = 278.7°, I = 57.9°, and a_{95}=3.2°, corresponds to a paleolatitude of

38.6° ± 3.5°N. A single zone of normal polarity cannot be correlated with the magnetic polarity timescale. The basal horizons of overlying sediments are of early Ypresian to late Ypresian-early Lutetian age, i.e., from ca. 56 Ma to 50 Ma. It was already stated that the Ar/Ar age for these basalts of ~40 Ma is in sharp contrast with the paleontological data. Since the early Ypresian age of the sediments above the basalts is strongly indicated by planktonic foraminifera and nannoplankton, we prefer to discard the single Ar/Ar determination. Tentatively, the age of 54.5 ± 1.5 Ma is ascribed to this paleomagnetic result, in accord with an inferred early Ypresian age of these volcanics. It should be mentioned that the Paleogene results are statistically identical and hence may be combined. The rather large difference in the rock ages, however, favors their separate consideration.

6. Paleomagnetic Data From Other Island Arc Complexes of Kamchatka

The East Peninsulas zone consists, from north to south, of Cape Kamchatka and the Kronotsky and Shipunsky Peninsulas on the Pacific side of Kamchatka (Figure 1b). Some authors regard this zone as the extinct Kronotsky island arc [*Zinkevich and Tsukanov*, 1992], whereas other authors consider the Cape Kamchatka block as the westernmost part of the Aleutians [*Geist and Scholl*, 1994]. Analysis of island arc complexes showed that variation in composition, ages, true thicknesses, etc., along the EPZ is comparable to that along the Aleutians [*Bazhenov et al.*, 1992]; hence available data do not discriminate between the above options. So we decided to analyze paleomagnetic data from Cape Kamchatka together with our results on the Kronotsky Peninsula.

Paleomagnetic data on the Paleogene island arc complexes from Cape Kamchatka (Figure 1b, localities R and Q) indicate that the Cape Kamchatka island arc terrane was at latitudes of 38.1° ± 4.1°N and 47.0° ± 6.4°N in early Paleocene (65-61 Ma) and late Eocene (46-43 Ma), respectively [*Pechersky et al.*, 1997]. Accumulation of the island arc series at Cape Kamchatka lasted from the beginning of the Paleocene up to early late Eocene time, i.e., for ~30 m.y. The age difference between the accumulation of the two studied formations is ~20-25 m.y. During this interval, the studied terrane moved northward by 8.9° ± 6.1°.

The Late Cretaceous island arc complex on the Kronotsky Peninsula originally belonged to an island arc which was active in the Late Cretaceous. Widespread Late Cretaceous island arc complexes of similar age in the Achayvayam-Valagina zone of Kamchatka are regarded as the remnants of another island arc (arcs) which was also active in the Late Cretaceous. Therefore, while trying to reconstruct the kinematic evolution of the Kronotsky island arc, we cannot ignore the other arc and the significant paleomagnetic data set for its Late Cretaceous island arc complexes and overlying continent-derived Paleogene rocks (Figure 1b and Table 3).

The paleomagnetic results based on fully demagnetized data are available from the following localities given on Figure 1b: A, the northern part of the Median Range [*Levashova et al.*, 1998]; B, the Eastern Ranges (Kumroch Range) [*Levashova et al.*, 1997]; C, Karaginsky Island [*Kovalenko et al.*, 2000]; D-L, the Olutor zone of the Koryak Highland (D-G are Late Cretaceous island arc complexes and H-L are Paleogene sediments) [*Kovalenko*, 1992, 1993]. All of these results are supported by fold tests, the results from localities D, E, F, H, and I are also supported by reversal tests, and the result from locality B is supported by a conglomerate test.

Levashova et al. [1998] analyzed the above results and showed that the studied complexes were originally parts of the same island arc (AVIA) which originated 20°-25° south of the continental margin in the Campanian and then was transported northward until its emplacement in the Paleocene-middle Eocene. The AVIA was active from the beginning of the Campanian until the early Paleocene (ca. 83 to 65-60 Ma). Kinematic analysis, based on geological and paleomagnetic data from the Olutor-Kamchatka region, showed that a coastwise translation can account for only a minor part of the total northward displacement of the AVIA, and this arc drifted northwestward on the leading edge of either the Kula or the Pacific plate. The motion with the Kula plate predicts the docking time of ~65-60 Ma, whereas the motion with the Pacific plate predicts the arrival time of ~55-50 Ma. According to geological data, the docking time for the AVIA is limited to the interval 65-45 Ma; we therefore cannot discriminate between these two scenarios.

7. Interpretation and Discussion

To reconstruct the kinematic history for the Kronotsky island arc, one must know the kinematic parameters of the main continental and oceanic plates for Late Cretaceous-Cenozoic time, the interval of the arc's volcanic activity, the age of collision between the continental margin and the island arc, and the paleolatitudes of the paleoarc at different times. The paleolatitudes were derived from our paleomagnetic data on the Kronotsky Peninsula (54.5°N, 162°E) and the results from Cape Kamchatka (56°N, 162°E) [*Pechersky et al.*, 1997]. The method of backward modeling [*Debiche et al.*, 1987] was used to calculate the trajectories of the Cape Kamchatka and Kronotsky terranes of the Kronotsky island arc. In this study, all calculations were made in the hotspot reference frame; the kinematic parameters of the main oceanic and continental plates are from *Engebretson et al.* [1985].

It is not definitely known where the boundary between the Eurasian and North American plates was located in Late Cretaceous and Cenozoic time; hence it is unclear to what continental plate the remnants of the Kronotsky paleoarc had docked. The calculations were made for each of the continental plates as a docking plate, and the difference in kinematic trajectories for these two options proved to be negligible (Figure 7a). The Olutor-Kamchatka region is more likely to be a part of the North American plate now [*Chapman and Solomon*, 1976; *Cook et al.*, 1986; *Riegel et al.*, 1993]. We hypothesized that it was a part of the North American plate in Cenozoic too, and the following discussion will be limited to this case.

Assuming that the Kronotsky island arc was moving with the North American plate, the expected positions of the studied terranes in Late Cretaceous and Paleogene time were calculated. For both terranes the observed Cretaceous, Paleocene and Eocene paleolatitudes are considerably lower than the expected ones (Table 3). Thus the northward transport of the Kronotsky island arc is strongly indicated by the paleomagnetic data. It worth noting that tropical and subtropical forms of planktonic foraminifera, such as *Subbotina bowery*, *S. galavasi*, *S. tripartita*, and *Acarinina bullbrooki*, are found in the Paleogene island arc complex of the Kronotsky Peninsula. These forms are usually supposed to live at latitudes not higher than 48°-50°; hence paleontological data are consistent with the northward drift of the Kronotsky paleoarc.

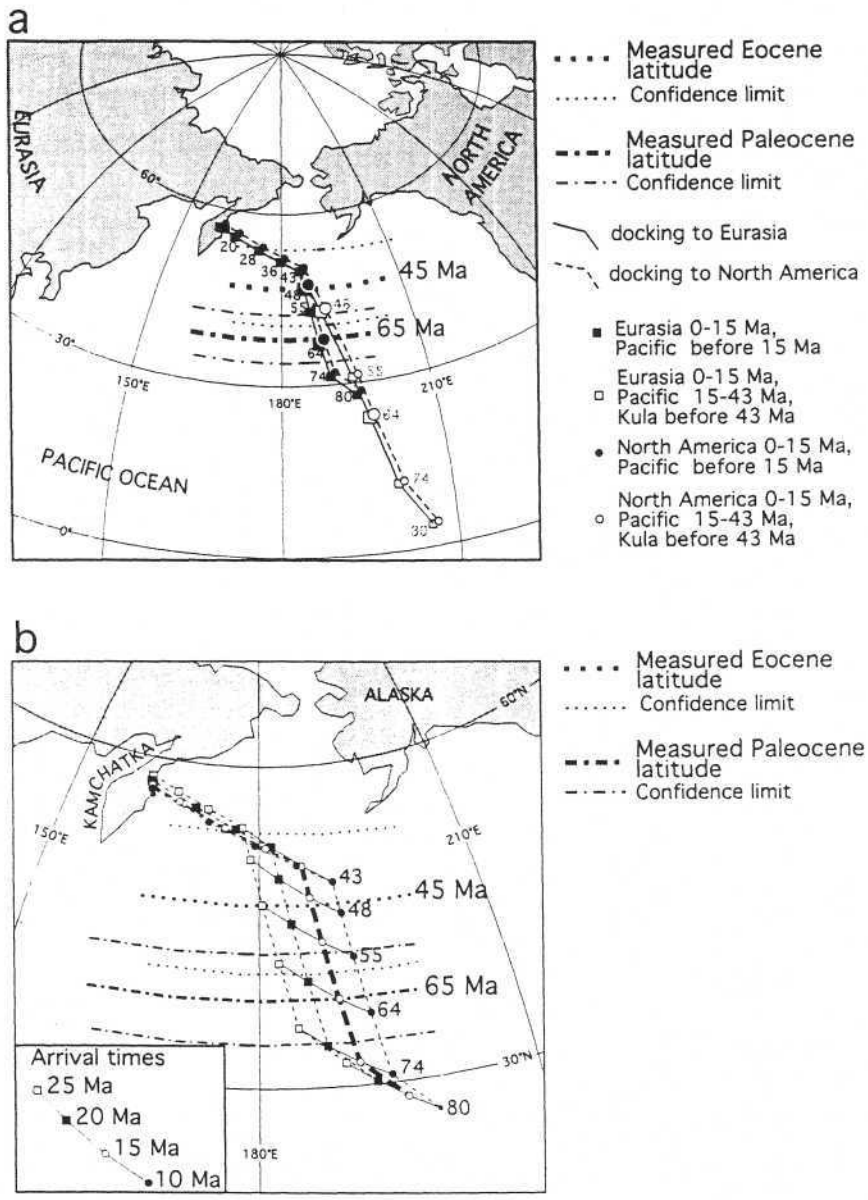

Figure 7. Kinematic trajectories of the Cape Kamchatka terrane. (a) Comparison of the trajectories for the Pacific (solid symbols) and Kula (open symbols) plates as carrier plates and 15 Ma arrival time to either Eurasia (squares) or North America (dots). (b) Comparison of the trajectories of the Cape Kamchatka terrane for its docking to North America for different arrival times. The trajectories are backward modeled [*Debiche et al.*, 1987] using the kinematic parameters of *Engebretson et al.* [1985] in the hotspot reference frame. The ages of points on the trajectories are in million years in black for the Pacific plate and shaded for the Kula plate. Eurasia and North America are shown in their present-day positions

The Kronotsky island arc could have moved with an oceanic plate (or plates) before collision with the continental margin and/or along strike-slip faults after collision. Geochemically, the volcanoclastic complexes of the Kronotsky island arc are similar to volcanics of intraoceanic island arcs [*Khubunaya*, 1987; *Puzankov*, 1994], and the continent-derived sediments are absent in the island arc complexes of the EPZ. At ~43 Ma, the Kula-Pacific spreading center ceased and, since that time, the Pacific plate was moving subnormal to the continental margin [*Engebretson et al.*, 1985], thus excluding the possibility of a coastwise translation [e.g., *Beck*, 1991]. On the other hand, the observed Bartonian (40 ± 2 Ma) paleolatitude for the Kronotsky terrane and the Lutetian (44.5 ± 1.5 Ma) paleolatitude for the Cape Kamchatka terrane differ from the North American reference values by ~14° and 12°, respectively (Table 3). Hence the Kronotsky island arc was far away from a continent even in the Bartonian, let alone older times, and any coastwise transport can be excluded. Therefore this arc could move only with the Pacific plate after 43 Ma and with

Table 3. Paleomagnetic Data on the AVIA and Kronotsky paleoarcs

N	RA	Locality		Observed directions				Lat_{exp}	Lat_{obs}	ΔLat	Source
		Lat, °N	Lon, °E	*D*	*I*	*k*	a_{95}				
		Achayvayam-Valagina Island Arc (AVIA): Late Cretaceous									
A	Cmp (81±2 Ma)	59.3	162.1	285	67	25	4.1	70-72	49.7±5.6	20-22	1
B	Cmp (81±2 Ma)	56.5	162.0	340	68	16	3.7	68-70	48.7±5.0	19-21	2
C	Cmp-Maa (75±8 Ma)	59.0	164.4	332	63	20	4.5	69-73	44.5±5.6	25-29	6
D	Cmp-Maa (75±8 Ma)	59.8	164.9	299	61	18	6	70-74	42.1±7.1	29-32	5
E	Cmp-Maa (75±8 Ma)	61.4	171.5	80	66	15	8	72-75	49.0±10.9	23-26	4
F	Cmp-Maa (75±8 Ma)	60.9	171.5	88	65	14	5	72-75	47.0±6.5	25-28	4
G	Cmp-Maa (75±8 Ma)	60.9	171.7	97	68	12	5	72-75	51.1±7.0	21-24	4
		Achayvayam-Valagina Island Arc (AVIA): Paleogene									
H	Tha (58±2 Ma)	59.8	164.9	285	75	12	8	67-68	61.8±13.2	5-6	5
I	Ypr (53±3 Ma)	59.8	164.9	321	76	14	6	65-66	63.5±10.2	2-3	5
J	Lut-Brt (44±6 Ma)	59.8	164.9	299	73	10	10	63-65	58.6±15.8	4-6	5
K	Prb-Rup (34±5 Ma)	59.8	164.9	338	80	10	9	62-63	70.6±16.3	-8-9	5
L	Te (45±10 Ma)	60.6	168.0	348	74	17	4.2	64-67	60.2±6.8	4-7	3
		Kronotsky Island Arc									
M,N	Cmp-Maa (73±7 Ma)	54.5	162.0	283	63	43	6.4	66-68	44.8±8.0	23	8
O	Ypr (54.5±1.5 Ma)	54.5	162.0	279	58	489	3.2	62	38.6±3.5	23	8
P	Brt (40±2 Ma)	54.5	162.0	296	63	78	5.6	59	45.1±7.0	14	8
Q	Dan (63±2 Ma)	56.2	162.3	-	57	65	3.8	65	38.1±4.1	27	7
R	Lut (44.5±1.5 Ma)	56.2	162.3	20	65	59	4.9	59	47.0±6.4	12	7

N is the sampling locality labeled as in the text and Figure 1. RA is the rock age: Cmp-Maa, Campanian to Maastrichtian; Dan, Danian; Tha, Thanetian; Ypr, Ypresian; Lut-Brt, Lutetian to Bartonian; Prb-Rup, Priabonian to Rupetian. Locality coordinates: Lat, latitude, Lon, longitude. Observed data are in stratigraphic coordinates: *D*, declination (in degrees); *I*, inclination (in degrees); *k*, Fisher concentration parameter; a°_{95}, radius of 95% confidence circle (in degrees); Lat°_{exp} and Lat°_{obs}, the expected and observed paleolatitudes of the studied localities, respectively (in degrees); ΔLat°, the difference between the expected and observed paleolatitudes (in degrees). Source: 1, Median Range, *Levashova et al.* [1998]; 2, Kumroch Range, *Levashova et al.* [1997]; 3, Malinovsky Range, *Kovalenko* [1992]; 4, Olutor zone, *Kovalenko* [1992]; 5, Ilpinsky Peninsula, *Kovalenko* [1993]; 6, Karaginsky island, *Kovalenko et al.* [2000]; 7, Cape Kamchatka Peninsula, *Pechersky et al.* [1997]; 8, this study.

either the Pacific or Kula plates before 43 Ma. Therefore two trajectories were calculated for each docking time.

Docking of the Kronotsky arc at 10-5 Ma may be inferred from major angular unconformity between flat-lying Pliocene volcanics and folded marine lower-middle Miocene sediments in the Tushev basin (Figures 1b and 1c) [*Shapiro et al.*, 1996]. However, a weak angular unconformity at the base of Oligocene marine sediments in the eastern limb of the Tushev basin [*Stupin et al.*, 2000] may indicate ancient docking. Still other view is expressed by *Zinkevich and Tzukanov* [1992], who relate middle Eocene deformation, which affected most of Kamchatka, to collision between the Kronotsky island arc and Kamchatka mainland. Thus the docking time for the Kronotsky island arc cannot be uniquely defined by the age of collision-related folding and thrusting, and only an interval of time when the docking took place can be proposed.

The upper limit on the age of collision is put by unconformable overlap of flat-lying Pliocene (5-10 Ma) volcanics on the folds and thrusts in the Tushev basin. On the other hand, the collision is unlikely to predate the youngest members of the Kronotsky island arc complex; so far, all reported ages on this complex are not younger than the late Bartonian-early Priabonian (40-36 Ma). Thus the docking time for the studied island arc is most likely to be between 10-5 and 40-36 Ma. However, the Cape Kamchatka terrane in the Lutetian and the Kronotsky terrane in the Bartonian were ~12° and 14°, respectively, to the south of the North American margin (Table 3) and could not have reached the continental margin earlier than 25-30 Ma. According to this reasoning, the kinematic modeling of the Kronotsky island arc was made for docking times varying from 25 to 5 Ma.

The Cape Kamchatka terrane was located at a latitude 38.1° ± 4.1°N in the early Paleocene (65-61 Ma) and at a latitude 47.0° ± 6.4°N in the Lutetian (46-43 Ma) [*Pechersky et al.*, 1997]. The points of the same ages on the trajectories for the Kula as a carrier plate differ much more than the observed paleolatitudes. Besides, the expected Paleocene paleolatitude on the Kula trajectories is much lower than the observed value (Figure 7a). Thus we conclude that the Cape Kamchatka terrane could move only with the Pacific plate and that the best fit of the observed and expected paleolatitudes is achieved for 10-15 Ma docking times (Figure 7b).

Similarly, the motion with the Kula plate predicts lower original latitudes than the observed ones and larger differences between the Bartonian and Ypresian paleolatitudes (Figure 1, localities O and P) for the Kronotsky terrane (Figure 8). In contrast, Paleogene results fit, within the error limits, with the kinematics of the Pacific plate for docking times of ~10 Ma, in accord with the Paleocene and Eocene results from Cape Kamchatka. Thus we assume that the Cape Kamchatka and Kronotsky terranes belonged to the Kronotsky island arc and were moving with the Pacific plate since early Paleocene time (65-61 Ma) until docking at ~10 Ma. This scenario fits the available paleomagnetic and kinematic data within their uncertainties and agrees with late Miocene-Pliocene age of deformation and thrusting in the Tushev basin.

Subduction-related volcanism in the Kronotsky island arc and

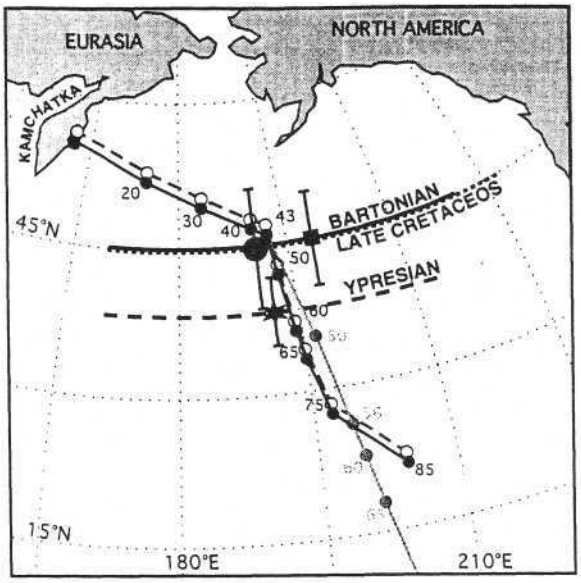

Figure 8. Kinematic trajectories of the Kronotsky terrane for motion with the Pacific plate (solid dots and line) and first with the Kula plate, then with the Pacific plate (shaded dots and line) for 6 Ma arrival time to North America. Late Cretaceous, Ypresian, and Bartonian observed latitudes are shown as large solid circle and thick solid line, asterisk and dashed line, and star and dotted line, respectively; error bars are given for each symbol. For comparison, the trajectory for the Cape Kamchatka terrane is shown as open dots and thin dashed line.

its motion with the Pacific plate occurred simultaneously during the Paleocene-Eocene. This implies that the arc was on the leading edge of the Pacific plate and that another plate was being subducted. We assume that an oceanic periphery of a continent, more likely of North America, was being subducted as, for instance, the periphery of the Australian plate is now subducting in the Sundae trench. After the cessation of volcanic activity and the extinction of the subduction zone related to the Kronotsky island arc at ~40-36 Ma, the extinct Kronotsky island arc kept moving with the Pacific plate until its collision with the continental margin in the late Miocene-Pliocene. Simultaneously, a new subduction zone responsible for the Oligocene-Miocene volcanism in Kamchatka (Figure 1b) started developing at the Eurasian boundary. The Vetlov zone likely represents an accretionary complex of this subduction zone. We assume that the collision of the Kronotsky island arc produced not only the deformation in the Tushev basin but also blocked the Oligocene-Miocene subduction zone near Kamchatka. Volcanism in central Kamchatka was gradually receding in the late Miocene-Pliocene and a new subduction zone, which is still active, started developing somewhat to the east from the Oligocene-Miocene one, as shown by the shift of volcanic activity from central to east Kamchatka (Figure 1b).

While the Cenozoic kinematics of the Kronotsky island arc seem clear, the Late Cretaceous situation is more obscure. According to the available paleomagnetic data, the Kronotsky terrane was at 38.6° ± 3.5°N in Ypresian time (54.5 ± 1.5 Ma), and the Cape Kamchatka terrane was at 38.1° ± 4.1°N in early Paleocene time (63 ± 2 Ma) (Table 3). However, the Kronotsky terrane was at a higher latitude of 44.8° ± 8.0°N in the late Campanian-Maastrichtian (73 ± 7 Ma). The Late Cretaceous latitude differs by 6.2° ± 7.0° and 6.7° ± 7.2° from the Ypresian and early Paleocene values, respectively, both values being statistically insignificant. On the other hand, had the Kronotsky terrane continued moving with either the Pacific or Kula plate, its paleolatitude for 73 Ma would have been of ~30°N and 15°N, respectively (Figure 8), in sharp contrast to the observed value. Thus, in the pre-Paleocene time the Kronotsky island arc definitely did not move quickly to the north; instead, it was probably slowly moving southward. Therefore this arc could not move with either the Kula or Pacific plate, which had been steadily and quickly moving north to northwestward.

One solution is to assume that the Kronotsky island arc resided on an unknown oceanic plate. Another solution is to assume that the Kronotsky island arc was on the oceanic periphery of a slowly moving continental plate, either Eurasia or North America. This implies that the paleoarc had belonged to the same plate as the continent but was separated from the continent by a large basin with oceanic floor, like the Aleutians and North America. Of these two hypotheses, the second one seems to be preferable to us, simply because it is more difficult to justify the existence of a plate which has disappeared completely.

In Campanian-Maastrichtian time and probably in Early Paleocene time (ca. 83 to 65-60 Ma), while the Kronotsky island arc resided on the oceanic periphery of a continent, the AVIA was moving northwestward across the North Pacific with either the Kula or Pacific plate [*Levashova et al.*, 1998]. It is reasonable therefore to analyze the pre-Paleogene evolution of the Kronotsky island arc and AVIA together. As a result, we have obtained two different scenarios.

7.1. Scenario 1

7.1.1. Campanian-Maastrichtian. In the early Campanian (83-79 Ma) the intraoceanic Achayvayam-Valagina island arc originated in the northwest Pacific ~2000-2500 km to the south of the continental margin. It moved northward on the leading edge of the Kula plate, and the related subduction zone was consuming the oceanic periphery of a continent. The Kronotsky island arc originated in Campanian-Maastrichtian (73 ± 7 Ma) approximately at the same latitudes as the AVIA. It was slowly moving southward with a continental plate, and the related subduction zone was consuming the Pacific plate. The AVIA and Kronotsky arc were separated by a transform fault that is a continuation of the Kula-Pacific transform boundary (Figure 9a).

7.1.2. Paleocene (ca. 65-55 Ma). The AVIA had arrived at the continental margin, its activity ceased, and the overlying continent-derived flysch started accumulating. Because the related subduction zone was blocked, the AVIA no longer marked a plate boundary, and its fragments accreted to the Achayvayam-Valagina zone of the Olutor-Kamchatka region. In the first half of the Paleocene (ca. 65-60 Ma) the subduction polarity of the Kronotsky island arc was reversed; the arc changed its carrier-plate and started moving on the leading edge of the Pacific plate (Figure 9b).

7.1.3. Paleocene - Bartonian (ca. 65-40 Ma). The Kronotsky island arc moved on the leading edge of the Pacific plate, and the related subduction zone was consuming the oceanic periphery of a continent (Figure 9c).

7.1.4. From 40-36 Ma to 10 Ma. Volcanic activity of the Kronotsky island arc ceased at ~40-36 Ma, and a new subduction zone responsible for the Oligocene-Miocene volcanism in central Kamchatka started developing at the continental margin. The accretionary complex of the Vetlov zone was forming between this

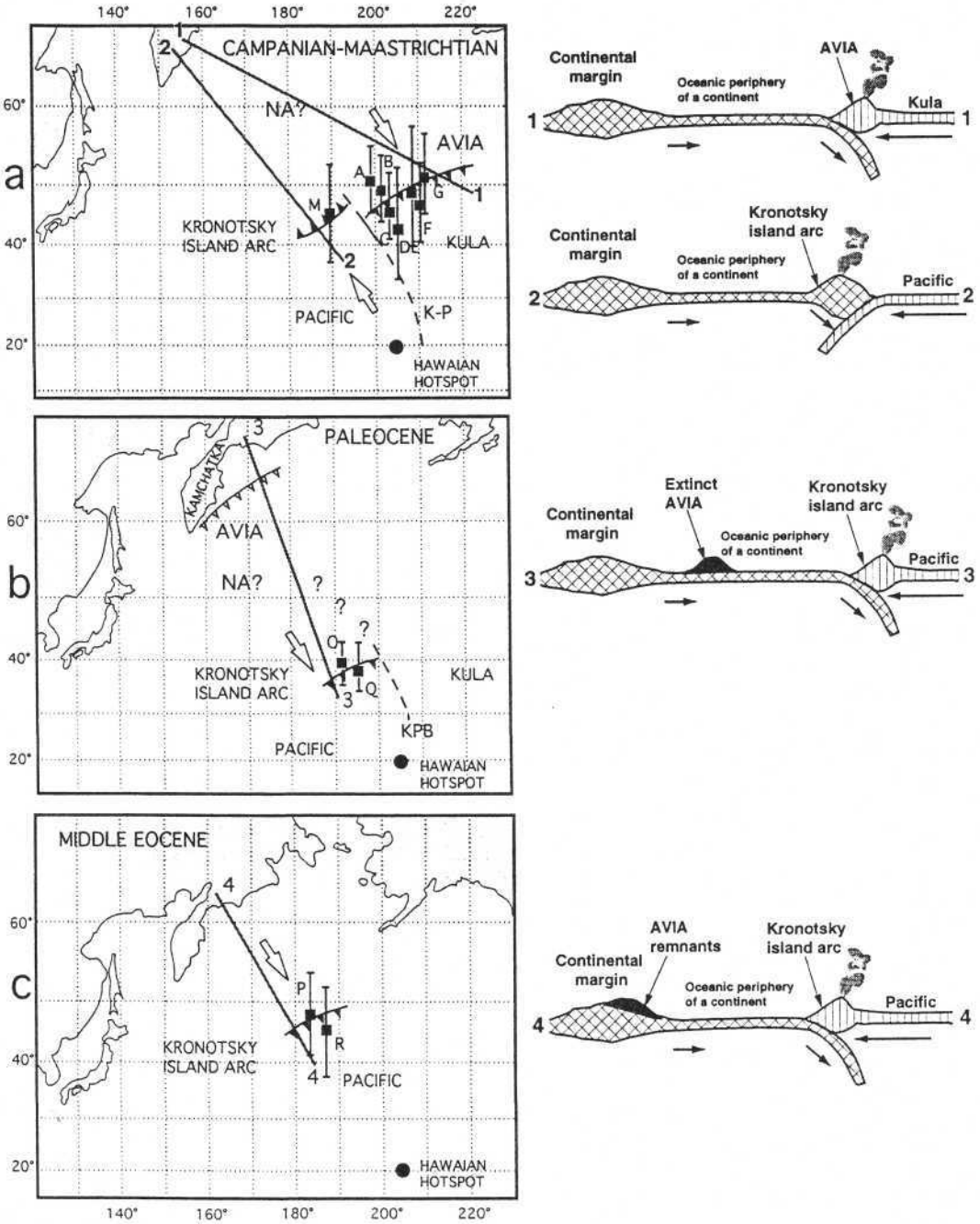

Figure 9. Positions of the island arcs related to the island arc terranes of Kamchatka in the North Pacific (a) in the Campanian-Maastrichtian (ca. 80 Ma), (b) Paleocene-early Eocene (ca. 60 Ma), and (c) middle Eocene (40-45 Ma) within the framework of scenario 1 (see text for explanations). Squares denote observed paleolatitudes with associated error bars labeled as in the text and Table 3. Large open arrows are parallel to the motion of subducting plates. Toothed lines with solid (open) teeth are active (extinct) subduction zones. Out-of-scale schematic profiles (right) illustrate inferred relationship of the plates and island arcs.

new subduction zone and the Kamchatka mainland. The extinct Kronotsky island arc moved with the Pacific plate until its collision with the continent. This collision led to Miocene-Pliocene folding and thrusting in the Late Cenozoic Tushev basin (Figure 1) and also blocked the Oligocene-Miocene subduction zone. Volcanism in central Kamchatka was gradually waning since that time, and the present-day subduction zone started forming in east Kamchatka (Figure 1b).

7.2. Scenario 2

7.2.1. Campanian-Maastrichtian. The Kronotsky island arc was slowly moving southward with the continental plate, and the related subduction zone was consuming oceanic crust of either the Kula or Pacific plate. The AVIA was moving northward with the Pacific plate, and the related subduction zone was consuming the oceanic periphery of a continent (Figure 10a).

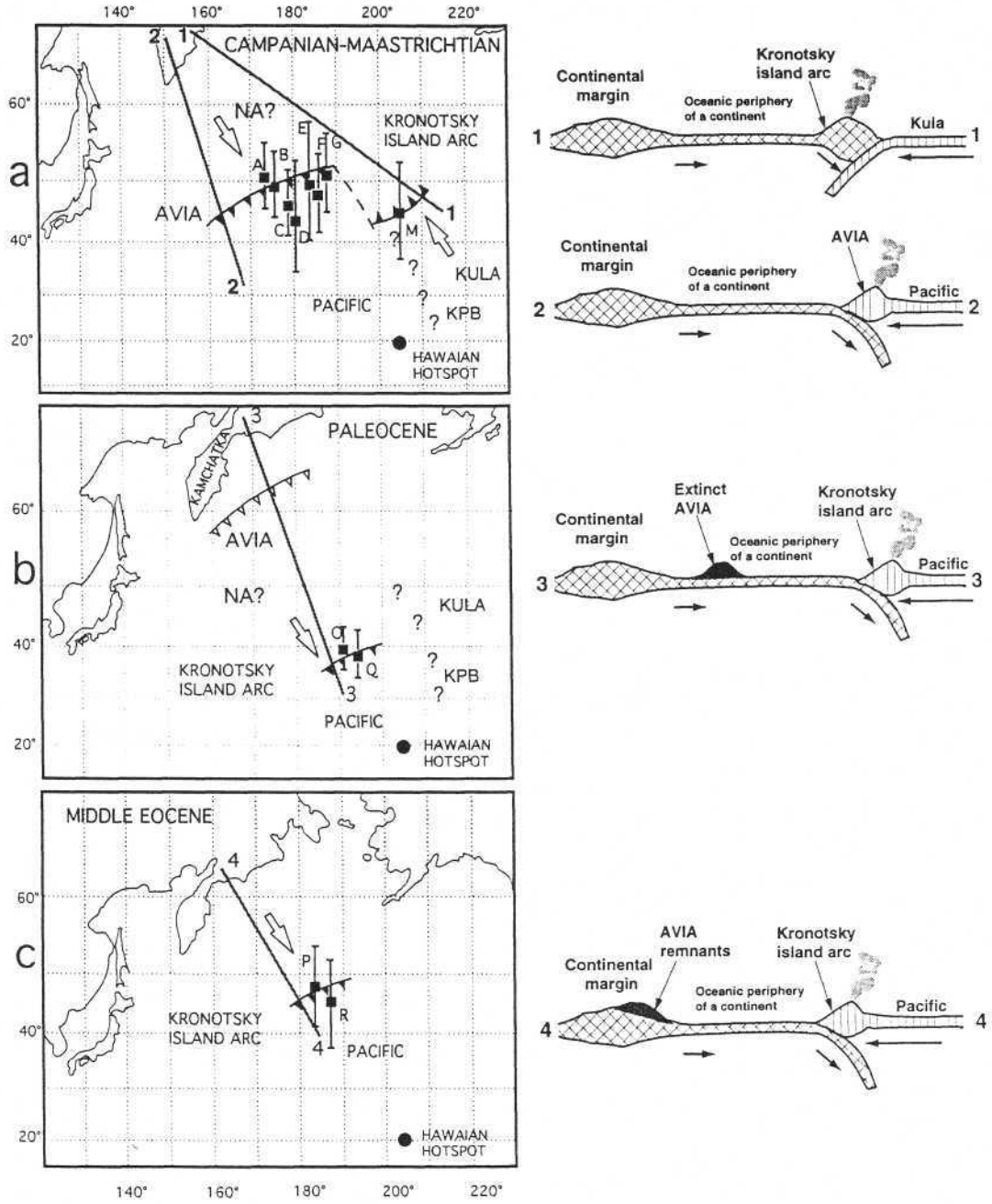

Figure 10. Positions of the island arcs related to the island arc terranes of Kamchatka in the North Pacific (a) in the Campanian-Maastrichtian (ca. 80 Ma), (b) Paleocene-early Eocene (ca. 60 Ma), and (c) middle Eocene (40-45 Ma) within the framework of scenario 2 (see text for explanations). Other notation is as in Figure 9.

7.2.2. 65-60 Ma. The volcanic activity was ending in the AVIA, which was already close to the continental margin, but the collision itself had not yet begun. The subduction zone related to the Kronotsky island arc had changed its polarity and started moving with the Pacific plate, consuming the oceanic periphery of the continental plate (Figure 10b).

7.2.3. 55-50 Ma. The AVIA collision with the Kamchatka mainland had started. The Kronotsky island arc continued its motion with the Pacific plate.

Since 50 Ma, the two scenarios coincide (Figures 9c and 10c). In both the AVIA had already reached the continental margin, and the Kronotsky island arc kept moving with the Pacific plate until its collision with the continent. Both scenarios assume the reversal of subduction of the Kronotsky subduction zone by the end of Cretaceous. We do not know what had caused the reversal and how to justify this event. Note, however, that the Late Cretaceous rocks are exposed only in the very limited part of the Kronotsky Peninsula, and the contacts between the Late Cretaceous and Paleogene rocks are tectonic everywhere. The Late Cretaceous rocks are also more deformed than in the other parts of the peninsula. Still, it remains unclear whether these observations indicate that the deformation was related to the reversal of subduction polarity.

Our analysis was performed in the hotspot reference frame using kinematic parameters of *Engebretson et al.* [1985]. Since then, several revisions of the Pacific kinematics appeared. When we used the kinematics of *Lonsdale* [1988], the difference between the Kula and Pacific trajectories became less, but the general fit of the Paleogene latitudes for the Kronotsky arc is noticeably better for motion with the Pacific plate than with the Kula plate. In particular, motion with the Kula plate predicts paleolatitudes too low for the Kronotsky arc in the Paleocene and Ypresian. Thus the main parameters of the Kronotsky arc kinematics are robust. In contrast, the Pacific kinematics of *Norton* [1995] predict more northerly latitudes for the Kronotsky island arc for any reasonable docking times and also places this arc very far to the east, leaving no space for the Kula plate.

The two scenarios above fit the available kinematic, geological, and paleomagnetic data, but we cannot discriminate them because we do not know whether the AVIA moved with the Kula or Pacific plate. Nevertheless, an indirect reasoning favors the second one. Campanian-Maastrichtian island arc volcanics are common on Sakhalin, East Hokkaido (Nemuro Peninsula), and the Lesser Kuril Islands (Figure 1a), whereas Paleogene subduction-related volcanism is nearly absent. In contrast, Late Cretaceous island arc complexes are scarce in east Kamchatka and the Aleutians. The only exception is a small part on the Kronotsky Peninsula, whereas Paleogene subduction-related volcanism is widespread. All these Late Cretaceous and Paleogene island arc complexes are the remnants of several unrelated island arcs in scenario 1. In contrast, scenario 2 allows for a connection between the Sakhalin-Hokkaido-Lesser Kuril volcanism and the AVIA in the Late Cretaceous on one hand and the Aleutian and Kronotsky island arcs in the Tertiary on the other. Moreover, the Paleocene-Eocene latitude of 43° ± 7°N for the western Aleutians (Komandorsky Islands) indicates considerable northward displacement of the western Aleutians too [*Bazhenov et al.*, 1992]. Thus we may hypothesize that the East Peninsulas zone and at least the western part of the Aleutians belonged to the same island arc in the early Tertiary. This interpretation would account for an apparent disagreement when some authors regard the Cape Kamchatka terrane as a part of the Aleutians [*Geist and Scholl*, 1994], whereas others [*Zinkevich and Tsukanov*, 1992] favor its association with the Kronotsky arc.

Note also that the observed declinations of all ages from the Kronotsky Peninsula are strongly deflected westward (Tables 1 and 2). The internal consistency of these data indicates that a considerable counterclockwise vertical axis rotation took place after accumulation of the island arc complex. As a post-Bartonian coastwise translation is shown to be negligible, this rotation most likely occurred during terrane collision with the Kamchatka mainland. At the same time, Paleocene and Eocene declinations are northerly on Cape Kamchatka [*Pechersky et al.*, 1997], whereas Paleocene-Eocene declinations on the Komandorsky Islands are strongly deflected eastward [*Bazhenov et al.*, 1992]. These rotations were accounted for by collision and bending of an originally rectilinear structure, a Kronotsky-Aleutian island proto-arc [*Bazhenov et al.*, 1992]. Of course, the above relations are conjectural, but further analysis of this problem is beyond the scope of this paper.

Finally, we would like to point out that the motion of the Kronotsky island arc with the Pacific plate implies that the Kula-Pacific transform boundary was to the east of this island arc since 65-61 Ma. Thus paleomagnetic data lead not only to deciphering the tectonic evolution of Kamchatka but also to help constrain the kinematics of oceanic plates.

8. Summary

The paleomagnetic study of Late Cretaceous and Paleogene volcanic rocks of island arc affinity from the Kronotsky Peninsula on the Pacific coast of Kamchatka showed that these rocks retained their primary magnetizations as supported by the positive fold, reversal, and conglomerate tests. Mean inclinations of these remanences are 20°, on average, lower than the corresponding Eurasian or North American reference values and indicate a considerable northward displacement of the Kronotsky terrane in accord with earlier paleomagnetic results on the island arc complexes from the other parts of Kamchatka. We performed backward modeling in the hotspot reference frame using kinematic parameters of *Engebretson et al.* [1985] and found that a considerable coast-parallel transport of the Kronotsky terrane and other island arc terranes of Kamchatka may account only for a minor part of the observed difference in measured and reference paleolatitudes; hence an intraoceanic transport is indicated. Similar kinematics for the Cape Kamchatka and Kronotsky terranes infer that they were both originally parts of the Kronotsky island arc. We assume that this arc was slowly moving southward on the edge of the oceanic marginal part of a continental plate in the Late Cretaceous, likely the North American plate. At about the Cretaceous-Paleocene boundary the polarity of subduction was reversed, and consequently, in the Paleocene and Eocene the Kronotsky island arc moved on the leading edge of the Pacific plate, the oceanic periphery of a continent having been subducted. This arc became extinct by the end of the Eocene but continued moving on the Pacific plate until its collision with the Eurasian margin at ~10-6 Ma.

Our analysis of available paleomagnetic and geological data from the entire Kamchatka Peninsula has resulted in two alternate scenarios of convergent boundary kinematics in the northwest Pacific. The existence of two scenarios is due to the fact that we cannot determine whether the Achayvayam-Valagina island arc moved on either the Pacific or Kula plate. More reliable paleomagnetic data on precisely dated rocks are needed for further progress in our understanding of plate evolution in the Pacific basin.

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M.L. Bazhenov, V.N. Beniamovsky, and N.M. Levashova, Geological Institute, Russian Academy of Sciences, Pyzhevsky Lane, 7, Moscow 109017, Russia (palmag@online.ru)

Mikhail N. Shapiro, United Institute of Physics of the Earth, Russian Academy of Sciences, B. Gruzinskaia st., 10, Moscow 123810, Russia.

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