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Age of Terrigenous Rocks of Northeastern Karaginskiĭ Island (Eastern Kamchatka)

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Abstract—Terrigenous complexes of the Cenozoic accretionary prism are exposed in the northeast of Karaginskiĭ Island (eastern Kamchatka). Their age is debatable to the present day. The studied nannoplankton from mudstone and the fission-track dating of detrital zircon from sandstone elucidate age of terrigenous rocks. Aleuropelites yield two nannoplankton assemblages. One ranging in age from the Ypresian–Lutetian (Supra-Khynkhlonai sequence) to Bartonian–Oligocene (sedimentary melange) is considered as an *in situ* assemblage characterizing age of sediments. The other one consisting of Upper Cretaceous–Paleocene taxa is regarded as redeposited from older sediments of unknown provenance. According to fission-track dating, the youngest population of detrital zircons is close in age to the deposition time of the most rocks studied. Terrigenous rocks of the Karaginskiĭ Island tend to be younger southeastward off the axial zone. The flysch complex is dated back to the Bartonian–early Oligocene, whereas rocks of sedimentary melange range in age from the Oligocene to early Miocene according to the fission-track dating of detrital zircon. Younger ages of rock located closer to the trench are typical of accretionary prisms.

Key words: nannoplankton, fission-track dating, detrital zircon, accretionary prism, Karaginskiĭ Island, Kamchatka.

INTRODUCTION

The Karaginskiĭ Island represents a part of the Komandor Basin framing (Fig. 1). The island structure and the history of its formation is important for understanding of geodynamics of the youngest tectonic zones in northeast Asia and adjacent sea areas. According to available reconstructions (Chekhovich *et al.*, 1989; Chekhovich *et al.*, 1990; Sokolov, 1992; Shapiro, 1995; *Ob'yasnitel'naya zapiska…*, 2000; Solov'ev *et al.*, 2002), the continental margin structure has been formed by collision of a Cretaceous island arc with the northeastern Asia. As is supposed, the collision resulted in formation of a new subduction zone east of the island arc accreted to the continent, which consumed first the Pacific Plate and then minor plates of the Komandor Basin of the Bering Sea. The Karaginskiĭ accretionary prism, which is exposed at present in the eastern part of the island, formed above that subduction zone (Chekhovich *et al.*, 1989). The prism should be paragenetic to volcanic belts of the Kamchatka Peninsula and southern Koryak Upland (Bogdanov *et al.*, 1987; *Ob'yasnitel'naya zapiska…*, 2000), i.e., to the Eocene–Oligocene volcanics of the northwestern Govena Peninsula (Govena arc), middle–upper Miocene volcanics of the Penzhina Bay coast (Kinkil belt), and Miocene–Pliocene volcanics of the Median and Vetvei ridges (Central Kamchatka and Apuka–Vyvenka belts). One of the difficulties complicating reconstruction of these relationships is the lack of age data on rocks of the prism. This work presents the results of fission-track dating for detrital zircons from the prism sandstones and new data on nannoplankton from mudstones. Furthermore, age data for the tuffaceous–terrigenous rocks crowning the island-arc volcanics in the axial part of the island are considered in addition.

GEOLOGIC SETTING

Three major structural zones are distinguished in the Karaginskiĭ Island (Chekhovich *et al.*, 1990; Kravchenko-Berezhnoy and Nazimova, 1991). The axial zone is mainly composed of gabbro–ultramafic rocks of the ophiolite association and of the Late Cretaceous– Paleocene island-arc volcanic and sedimentary rocks formed far to the south of their present position according to paleomagnetic data (Kovalenko, 1990; Kovalenko *et al.*, 1999; Kovalenko and Kravchenko-Berezhnoy, 1999). Some researchers (Shapiro, 1995;

Kovalenko *et al.*, 1999) believe that the volcanic rocks, the upper part of which is distinguished as the Khynkhlonai Formation, belonged to an island arc that collided with the continent in the middle Lutetian (Solov'ev *et al.*, 2002). At its terminal Paleogene stage of evolution, the collided structure is known as the Govena arc (Chekhovich *et al.*, 1990).

In the northeastern part of the island, volcanics of the Khynkhlonai Formation are conformably overlain by a mudstone member with tuffite laminae, the age of which has not been exactly established. Some researchers (Shapiro and Petrina, 1984; Chekhovich *et al.*, 1990) attribute these rocks to the flysch complex characterized below. This is not undisputable however, because interlayers of sandstone (plagioclase + quartz + rhyolite) typical of the flysch are missing from this sequence, whereas tuff and tuffite, not characteristic of flysch, are abundant. Furthermore, their conformable occurrence above the island-arc volcanics is inconsistent with the sequence position in the accretionary prism. We separate it therefore from the flysch and designate arbitrarily as the Supra-Khynkhlonai sequence, the age of which determines the upper limit of the Govena arc active development. Previously, the sequence age was determined based on correlation with the flysch complex as ranging from the Maastrichtian–Paleocene to middle Eocene (Shapiro and Petrina, 1985; Chekhovich *et al.*, 1990).

The axial zone is bordered on northwest, mainly through faults, by a monocline composed of the middle Eocene–Pliocene, conformably bedded sediments, which dips in the northwestward direction, being complicated by gentle folds, and represents a slope of the Litke trough. Lower levels of the trough axial section are exposed on the Il'pin Peninsula (Volobueva *et al.*, 1994), where Upper Cretaceous volcanics are overlain with a minor unconformity by tuff–terrigenous rocks of the lower Paleocene. Another unconformity is established in the middle Miocene interval of overlying rocks only (Gladenkov, 1972; Gladenkov *et al.*, 1998). Faults bounding the Litke trough of the Karaginskiĭ Island are of a small amplitude. In places, Miocene and Pliocene sedimentary sequences occur with unconformity on rocks of the island axial zone.

According to Chekhovich *et al.* (1989, 1990), the southeastern part of the island is composed of the flysch and sedimentary melange complexes. *The flysch complex* exhibits the variably ordered alternation of mudstones and sandstones, some of which represent typical turbidites. However, many mudstone members of this complex up to hundreds of meters thick are lacking sandstone interlayers. An important feature of flysch sandstones is prevalence in their composition of acid volcanic and, to a lesser extent, granite clasts over fragments of more basic volcanic rocks (Shapiro, 1984; Shapiro *et al.*, 2000). Characteristic of some mudstone members in the flysch are well-rounded disseminated pebbles of acid volcanic rocks (Shapiro, 1984; Shapiro *et al.*, 2000). Flysch and mudstone members often enclose lenticular intercalations of tholeiitic pillow lavas, which are regarded either as conformable lenses, or as olistoliths and tectonic detached masses of the oceanic crust (Chekhovich *et al.*, 1990; Kravchenko-Berezhnoy *et al.*, 1990). Beds of the flysch complex are intensely folded or form northwestward-dipping monoclines separated by imbricated thrust faults conformable with bedding in general but in places outlined by protrusions and lenses of serpentinite melange. The flysch complex is bordered by major, northwestward dipping reversed or thrust faults, which separate it from volcanics and ophiolites of the axial part.

The intensity of folding of the flysch complex increases in the southeastward direction, where the beds losing their coherence are dismembered into large or small, often rotated lenses. Among the latter, there appear large lenses of pebbly mudstones with disseminated pebbles and boulders of rhyolites and their tuffs. The transition to the *sedimentary melange complex* is marked by distinct lithological changes: sandstones are less frequent in the section, being replaced by siliceous mudstones, cherts, tuffites and tuffs, psephitic included. A kind of chaotic mixture of nearly isometric fragments of different rocks set in clay matrix is often characteristic of exposures, though coherent members of tuffs and cherts can also be observed. In places, lenses of pillow or massive (rarer) basalts are confined to tuff interlayers, while small basalt bodies often occur in chaotic members. In terms of geochemistry, basalts are frequently of the oceanic type, though the island-arc rocks also associate with them (Chekhovich *et al.*, 1990). Like in the flysch complex, rhyolite pebbles prevail among rock debris in pebbly mudstones. The thrust fault separating flysch from sedimentary melange complexes is traceable northeastward almost up to the shore near the Nizkii Cape, where it turns southwards, extending for 7 km and outlining antiform hinge, in the eastern flank of which the overturned basal planes of sandstone and tuff beds are facing westward (Fig. 1).

The age interpretation of rock complexes in the northeast of the Karaginskiĭ Island changed with time. In the first publication (Kharkevich, 1941), they were referred to as Mesozoic. Later on, the Miocene molluscan fauna was found in rocks of both complexes exposed along the southeastern coast of the island (Khramov *et al.*, 1969). In 1967, G.P. Borzunova who sampled sedimentary melange once more suggested the Oligocene age of its fauna. In opinion of Yu. B. Gladenkov (Chekhovich *et al.*, 1990, and private communication), the described mollusks are of the middle Eocene age. The Eocene age of mollusks collected from the flysch complex was also determined by V.M. Gladikova (Mel'nikova and Dolmatov, 1973). Afterward, agglutinated foraminifers from flysch and sedimentary melange complexes were referred to the Maastrichtian–Danian, similar to their assemblages from the Vetlov Formation of the Kumroch Ridge (Shapiro and Petrina, 1985). At present, the formation is attributed to the

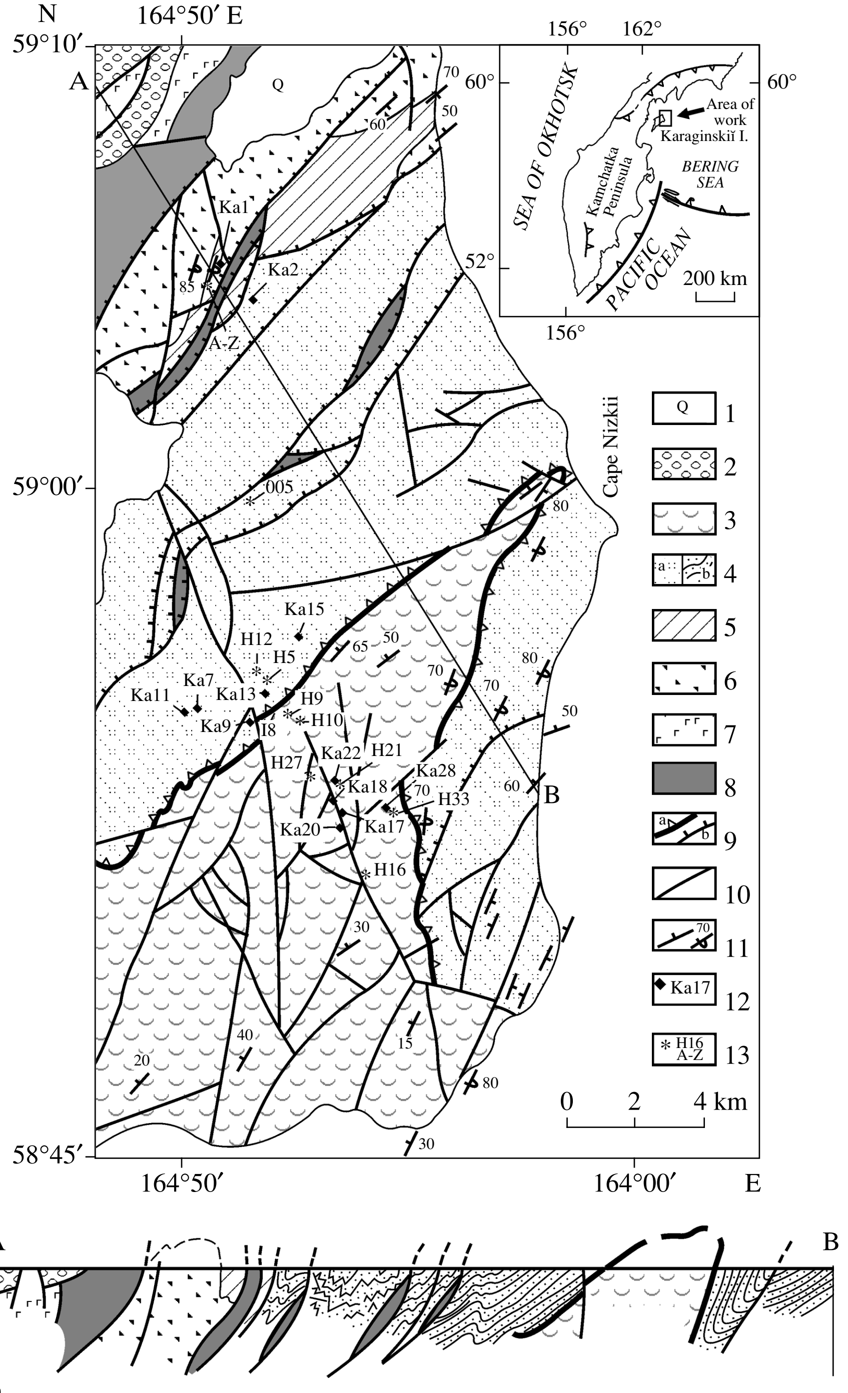

Fig. 1. Geological structure of the northeastern Karaginskiĭ Island and geological profile along line AB (modified after Shapiro and Petrina, 1985): (1) Quaternary deposits; (2) Neogene deposits; (3) sedimentary melange complex (Oligocene–Lower Miocene); (4) flysch complex (middle Eocene–Lower Oligocene) on scheme (a) and in profile (b); (5) Supra-Khynkhlonai sequence (Lower Eocene); (6) Khynkhlonai Formation of basaltic andesite and tuffs (Maastrichtian–Lower Eocene); (7) pre-Maastrichtian Cretaceous deposits; (8) gabbro-ultramafic rocks of ophiolite association; (9) thrust and reversed faults connected with flysch complex thrust over sedimentary melange complex (a) and others (b); (10) steeply dipping faults; (11) strike and dip; (12) sampling sites for fission-track dating; (13) sampling sites for nannoplankton analysis.

Paleogene–lowermost Eocene (Gladenkov *et al.*, 1998).

Recent age determinations for terrigenous rocks of the Karaginskiĭ accretionary prism are based primarily on radiolarians from chert intercalations in sedimentary melange and on nannoplankton from mudstones of the flysch complex. According to Chekhovich *et al.* (1990), cherts in sedimentary melange yield the middle Eocene radiolarians. The middle–late Eocene and likely associated early Oligocene forms are identified in nannoplankton from the flysch complex (Shcherbinina, 1997). Analogs of the flysch complex of the Govena Peninsula are attributed to the Eocene–Oligocene (Chekhovich *et al.*, 1990; Chamov, 1996). Hence, the sedimentary melange may be slightly older than the flysch complex. The dating results are based, however, on few different plankton assemblages and should be verified therefore.

FISSION-TRACK DATING OF DETRITAL ZIRCONS

The zircon fission-track dating (Wagner and Haute, 1992) is suitable for age determination of barren sequences and stratigraphic correlation. It also can be used for identification of terrigenous material provenances and in analysis of rocks exhumation processes (Garver *et al.*, 1999; Garver *et al.*, 2000a; Garver *et al.*, 2000b). Permitting age determination of separate mineral grains, the fission-track dating offers a possibility to separate zircon populations dissimilar in age and related to different provenances. The cooling of rocks in provenances is under control of different geological processes. On the one hand, volcanic rocks and near-surface intrusions get cool and eroded quickly, and their zircons quickly appear in the sedimentation basin thus being applicable for dating of barren terrigenous sequences (Brandon and Vance, 1992; Garver and Brandon, 1994; Garver *et al.*, 2000b; Solov'ev *et al.*, 2001; Shapiro *et al.*, 2001). On the other hand, the rock blocks rising up from deep levels are cooling (Wagner and Haute, 1992; Garver *et al.*, 1999) and cross at a certain moment the temperature level of the track system closure in zircons (215°–240°C, Brandon and Vance, 1992). Tracks formed and accumulated in zircon crystals since that moment can be used therefore to determine the age of rock cooling.

Ages of detrital zircons from terrigenous deposits usually vary greatly. For our purpose, it is necessary to establish the youngest population or the minimal age (Garver *et al.*, 1999) determining the lower time limit of sedimentation, as deposits are always younger than their constituent clasts. We used this approach for dating terrigenous deposits of the Karaginskiĭ Island.

Rock samples (6–8 kg in weight) were collected from the Supra-Khynkhlonai sequence (Sample Ka1, psammitic tephroid) and from flysch (samples Ka2, Ka7, Ka11, Ka13, Ka15, sandstones; Ka9, fine-pebbly conglomerates), matrix (samples Ka17, Ka22, Ka28, sandstones), and blocks (samples Ka18, Ka20) of sedimentary melange complex (Fig. 1, Table 1). Zircon grains were separated based on the standard technique in the Laboratory of Accessory Minerals at the Institute of Lithosphere of Marginal Seas, Russian Academy of Sciences (IL RAS). The zircon fission-track dating was carried out at the Union College (Schenectady, N. Y., USA) and the IL RAS. Some peculiarities of the method are described in the footnote to Table 1. From five to fifty zircon grains were dated per each sample. The analyzed samples yield widely ranging ages of individual zircon grains (Fig. 2). This suggests that after deposition the rocks have not been heated above the temperature (215°–240°C) of the track system closure in zircon (Brandon and Vance, 1992). The same is evident from lithology and structure of studied terrigenous deposits. They are lacking secondary minerals formed at the temperature exceeding 200°C and the cleavage formed under certain P/T conditions. Hence, fission-track dating of zircon grains from the sequences studied reflects their cooling times in provenance, and the sedimentary sequence is not older than the zircon grains it bears. The fission-track age of the youngest zircon population defines the lower limit of deposition time of host sandstones.

The age of the youngest population of zircon grains is timing sedimentation, if synchronous volcanism took place in the immediate vicinity during the sedimentation (Brandon and Vance, 1992; Garver and Brandon, 1994; Garver *et al.*, 2000; Solov'ev *et al.*, 2001; Shapiro *et al.*, 2001). In orogenic belts experiencing a speedy rise and erosion, the near-surface intrusions are quickly subjected to washout at the surface. Hence, the time span between zircon crystallization in intrusion and its appearance in sediments of the adjacent basin is not greater than few m.y. (Shapiro *et al.*, 2001).

One sample of psammitic tephroid (Ka1) was taken from an interlayer about 40 cm thick in the Supra-Khynkhlonai sequence. The rock is composed of dominant volcanic clasts, plagioclase, and insignificant admixture of pyroxene and ore minerals. Volcanic fragments are similar in composition but vary in structure from vitrophyric to lathy; light-colored microlitic rocks prevailing. Most zircon grains belong to population 50 Ma old, and their age corresponds to the boundary between the Ypresian and Lutetian (Table 1).

Five sandstone samples (Ka2, Ka7, Ka11, Ka13, Ka15) and one sample of fine-pebbly conglomerate (Ka9) were taken from the flysch complex. The composition of sandstones is common for flysch: feldspar, quartz, acid volcanics, and mudstones dominate clastic material. All the samples yielded two to three populations of detrital zircons, the youngest of which range in age from 45 to 30 Ma. The indicated time span corresponds to the second half of the Lutetian, Bartonian, late Eocene, and early Oligocene. The young zircon population (40 Ma) from fine-pebbly conglomerate, which is composed mainly of acid volcanics, is of the Bartonian age (Table 1).

Ages of zircons from rocks of sedimentary melange complex vary within a wide range. Sandstones similar in composition to sandstones of the flysch complex are composed predominantly of plagioclase and quartz (Ka17 and Ka28). Young zircons are not abundant in these rocks (10–20%), and their populations ranging in age between 18 and 24 Ma correspond to the early

Table 1. Fission-track ages of detrital zircon from terrigenous and tuffaceous rocks of the northeastern part of the Karaginskiĭ Island (eastern Kamchatka)

Sample no.	Group, formation	K	Age of zircon population		
			P1 (Ma)	P2 (Ma)	P3 (Ma)
Ka28	Melange	30	18.6 ± 3.5 (19%)	39.9 ± 3.1 (60.8)	88.0 ± 9.9 (20%)
Ka17	Melange	20	23.7 ± 3.9 (10%)	57.3 ± 5.2 (62%)	116.0 ± 23.9 (28%)
Ka22	Melange	40	26.1 ± 1.5 (57%)	56.0 ± 10.7 (18%)	96.4 ± 14.2 (25%)
Ka13	Flysch	35	29.9 ± 3.6 (47%)	54.7 ± 5.2 (53%)	
Ka15	Flysch	31	30.4 ± 1.6 (87%)		117.6 ± 14.8 (13%)
Ka11	Flysch	40	36.1 ± 2.4 (40%)	66.4 ± 5.9 (24%)	111.0 ± 8.4 (36%)
Ka9	Flysch	50	39.7 ± 2.6 (56%)	62.3 ± 8.9 (35%)	94.6 ± 39.0 (9%)
Ka7	Flysch	35	44.2 ± 2.6 (68%)	95.5 ± 7.3 (32%)	–
Ka2	Flysch	12	45.6 ± 4.6 (66%)	93.4 ± 14.5 (34%)	–
Ka1	Supra-Khynkhlonai sequence	30	50.2 ± 3.2 (93%)	–	188.3 ± 52.7 (7%)
Ka20	Block in melange (?)	35	70.1 ± 4.4 (80%)	97.5 ± 13.3 (20%)	–
Ka18	Block in melange (?)	5	82.1 ± 10.4 (100%)		

Note: K, amounts of dated zircon grains in sample. P1, P2, P3, zircon populations established on the bases of the program BinomFit v 1.8 (Brandon, 1992; Brandon, 1996). Ages are given in Ma, error corresponds to ±1σ, percentage in parentheses are calculated of the given zircon population relative to the total number of grains dated (K). Zircon grains were dated by the external detector method (Wagner and Haute, 1992), the procedure is described by Garver *et al.* (2000). Zircon grains were pressed into FEP Teflon^{MT} plates 2×2 cm^2 in size. Two plates were prepared for each sample. The plates were roughed out with emery cloth (800 grit) and then polished with diamond paste (9 μm and 1 μm), and Al_2O_3 paste (0.3 μm) at the final stage. Chemical etching was carried out in NaOH–KOH solution at the temperature of 228°C during 20 (the first plate) and 26 hours (the second plate). After etching, the plates covered with detector (mica with a low U content) were exposed to the thermal neutron flux of about 2×10^{15} neutron/cm^2 (thermal reactor of the University of Oregon) simultaneously with the zircon standard samples FCT (Fish Canyon Tuff) and BL (Buluk Tuff), and with glass-dosimeter with known U content (CN-5) (Hurford, 1998). The fission tracks are calculated under microscope Olympus BH-P with automated system and digital plotter (dry method, maximal magnification 1600). The ζ-factor (Hurford, 1998) calculated for age standards (FCT, BL) was found to be equal to 346.22 ± 9.57 (A.V. Solov'ev) and 355.03 ± 8.16 (J. Lederer).

Miocene. Gravelstone (Ka20) of a similar composition consists of quartz, feldspars, their aggregates, and of abundant chert clasts. The youngest population of detrital zircon from the rock is 70 Ma old (Maastrichtian). A peculiar tuffaceous sandstone (Ka22), the main components of which are large angular plagioclase grains and clasts of deformed mudstone, yielded the young zircon population 26 Ma old (late Oligocene). Lastly, zircons from lithoclast tuff (Ka18) with pyroxene admixture represent one population (Table 1) that is 82 Ma old (the earliest Campanian).

NANNOPLANKTON

We collected mudstone samples from the Supra-Khynkhlonai sequence, the flysch and sedimentary melange complexes for nannoplankton study and geochemical analysis aimed to establish the source of clay material. In total, four sample sets have been collected (Fig. 1, Tables 2, 3).

The first collection includes 29 samples (A–Z) of mudstones conformably overlying the Khynkhlonai Formation in upper courses of the Severnaya River. Mudstones enclose here thin tuff and tuffite interlayers. The tuffite Sample Ka1 was taken from this exposure for zircon dating. The ideally exposed section sampled from top to base is 120 m thick.

Sample F, the most representative, yielded one specimen of the late Campanian–early Maastrichtian *Reinchardtites levis* associated with smaller *Reticulofenestra* spp., *R. dictyoda, Chiasmolithus consuetus, Lophodolithus nascens*, and *Coccolithus pelagicus* characterizing altogether the late Ypresian–early Lutetian interval (52–45 Ma; Fig. 3). Due to predominance of Eocene species, the Late Cretaceous form is regarded as redeposited. Another six samples bear single nannoplankton forms. These are *Reticulofenestra* cf. *minuta* (Bartonian–early Oligocene) from Sample C, *Chiasmolithus* sp. (Paleocene–early Oligocene) from Sample J, *Chiasmolithus solitus* and *Reticulofenestra dictyoda* (late Ypresian–Bartonian) from Sample K, and *Coccolithus pelagicus* (Paleocene–present-day) from Sample S. Sample I yielded one specimen of *Micula staurophora* characteristic of the Coniacian–Maastrichtian interval. This form is likely redeposited. The Paleogene age of species from this series of samples is consistent with the early Ypresian–early Lutetian age of nannoplankton from Sample F.

The second collection (H1–12, 005–1/3, 4, 5) represents mudstones of the flysch complex. It characterizes

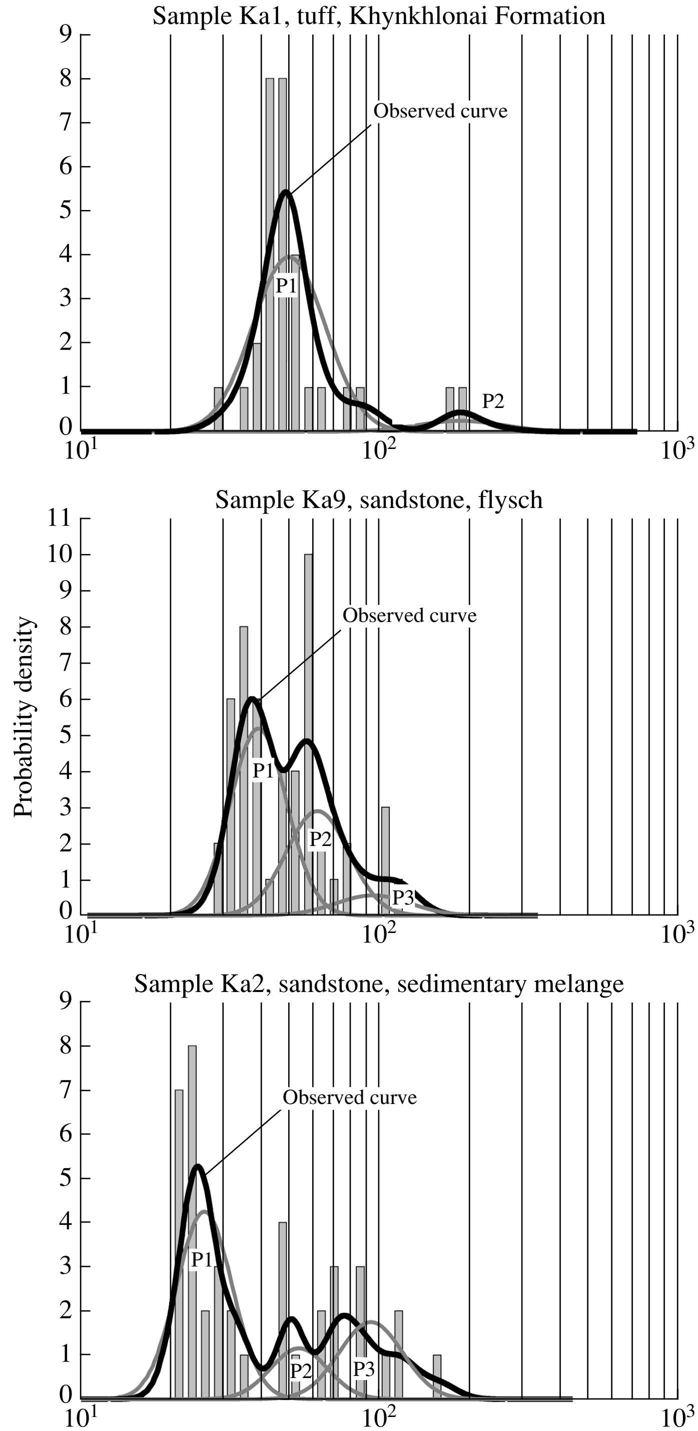

Fig. 2. Fission-track dating results for detrital zircons from tuffaceous–terrigenous deposits of the Karaginskiĭ Island (eastern Kamchatka); P1, P2, P3, peaks of different-age zircon populations (see Table 1) established using program BinomFit 1.8 (Brandon, 1996).

a profile across the flysch complex. Nannofossils found in seven of fifteen samples are exclusively the Mesozoic forms: Bajocian–Maastrichtian *Watznaueria barnesae* (005–1/5), Coniacian–Maastrichtian *Micula staurophora* (H5, 12), Campanian *Broinsonia parca constricta* and *Ahmuelerella regularis* (H8), Turonian–Campanian *Reinchardtites anthophorus* (H9), and Turonian–Santonian *Micula staurophora, Eprolithus floralis, Eiffellithus turriseiffeli, E. eximius, Watznaueria barnesae*, and *Retecapsa ficula*, (H10). Although

Table 2. *In situ* nannoplankton from terrigenous deposits of the northeastern part of the Karaginskiĭ Island (eastern Kamchatka); data of Shcherbinina (1997) included (see Fig. 3)

In situ forms	Khynkhlonai Formation	Flysch complex	Sedimentary melange complex
Lophodolithus nascens	+	–	–
Chiasmolithus consuetus	+	–	–
C. solitus (CP13, 14, Kronotskii Pen.)	+	–	–
Chiasmolithus sp.	+	–	–
Reticulofenestra dictyoda	+	–	–
Reticulofenestra cf. *minuta*	+	–	–
R. umbilicus (CP14, Baklanov Fm.)	–	+	+
R. hillae (CP14, Kronotskii Pen.)	–	+	–
R. oamaruensis	–	+/–	–
R. coenura (CP13, 14, Kronotskii Pen., Baklanov Fm.)	–	+	–
Dictyococcites bisectus (C14, Kronotskii Pen.; CP14–16, Govena Pen., Baklanov Fm.)	+/–	+	+
D. scrippsae (CP14, NW Govena Pen., Baklanov Fm.)	–	+	–
D. daviesii (CP14, Kronotskii Pen.)	–	+	–
Cyclicargolithus floridanus (CP13, 14, Kronotskii Pen., Baklanov Fm.)	–	+	+
C. abisectus (CP14, NW Govena Pen.)	–	+?	–
Discoaster deflandrei (CP14–16, NW Govena Pen., Baklanov Fm.)	–	+	–
D. saipanensis (index species of the CP14b Zone)	–	+	–
Isthmolithus recurvus (index species of the CP 15b Zone)	–	+	–
Sphenolithus moriformis (CP 11, 13, Kronotskii Pen.; CP 14, NW Govena Pen.)	–	+	+
S. predistentus	–	–	+
Coccolithus pelagicus			
C. floridanus	–	+	–
Cribrocentrum reticulatum	–	+	–

Table 3. Redeposited nannoplankton from terrigenous deposits of the northeastern part of the Karaginskiĭ Island (eastern Kamchatka), data of Shcherbinina (1997) included

Redeposited forms	Khynkhlonai Formation	Flysch complex	Sedimentary melange complex
Reinhardtites anthophorus	–	+	–
R. levis	+	–	–
Micula staurophora	+	+	+
Watznaueria barnesae	–	+	+
Broinsonia parca constricta	–	+	+
Ahmuelerella regularis	–	+	–
Eprolithus floralis	–	+	+
Eiffellithus turriseiffeli	–	+	+
E. eximius	–	+	+
Retecapsa ficula	–	+	–
Arkhangelskiella cymbiformis	–	–	+

our samples from the flysch complex yielded only Cretaceous forms, previously the early Eocene nannoplankton was also found in the same sequence (Shcherbinina, 1997). Moreover, the fission-track ages of youngest non-overheated zircon grains from sandstone and gravelstone of that sequence correspond to the middle Eocene–early Oligocene (46–30 Ma). The zircon dating discredits the pre-middle Eocene age of the sequence, and we believe the Cretaceous nannofossils from flysch samples are most likely redeposited.

The third collection (H13–33) characterizing aleuropelitic matrix of sedimentary melange complex includes only three samples with nannofossils (16, 27, and 33). They yielded the Coniacian–Maastrichtian *Watznaueria barnesae, Micula staurophora*, and *Eiffellithus turriseiffeli.*

The fourth collection (Ha21/ A–L) characterizing matrix of a pebbly mudstone lens in the sedimentary melange complex is sampled from different parts of a small (20×30 m) exposure on the watershed. Three of eleven samples yielded only Mesozoic nannofossils: Campanian *Micula staurophora, Arkhangelskiella cymbiformis, Broinsonia parca constricta* (I), Bajocian–Maastrichtian *Watznaueria barnesae* (K), and Cretaceous forms undeterminable at the species level (Z). One sample (V) yielded the Paleogene (Eocene–Oligocene) forms *Cyclicargolithus floridanus* and *Sphenolithus moriformis* (Fig. 3). The most interesting is Sample A, in which Cretaceous *Eiffellithus eximus* and *Eprolithus floralis* (Turonian–Santonian) are joined together in a pellet, whereas the clayey matrix bears *Reticulofenestra umbilicus, Coccolithus pelagicus, Dictyococcites bisectus*, and *Sphenolithus predistentus*, the assemblage of which characterizes the Bartonian–earliest Oligocene interval (41–28 Ma) (Fig. 3). It is apparent that this assemblage determines age of pebbly mudstones and thus of a substantial part of the sedimentary melange complex.

DISCUSSION

Redeposition of nannoplankton. The established two nannoplankton assemblages, one redeposited Late Cretaceous and the other one from the Eocene–Oligocene interval of the studied section represent the unexpected result of our study. In abundance and preservation state, the redeposited older forms are comparable with younger ones, and their occurrence frequency is even higher. For instance, exclusively Cretaceous forms are established in mudstone samples from flysch. Only the zircon fission-track dating of flysch sandstones and earlier found Eocene–Oligocene nannoplankton (Shcherbinina, 1997) make us consider the Late Cretaceous forms as redeposited.

In addition to dominating Eocene forms, Late Cretaceous nannofossils were also found in mudstones of the Supra-Khynkhlonai sequence. In one sample (F), Late Cretaceous species occur in association with Cenozoic forms. In Sample Ha21A, from matrix of pebbly mudstones in the sedimentary melange, the Turonian–Santonian species were also found along with Bartonian–early Oligocene forms. In the last case, Cretaceous forms were found inside a pellet of the biogenic origin.

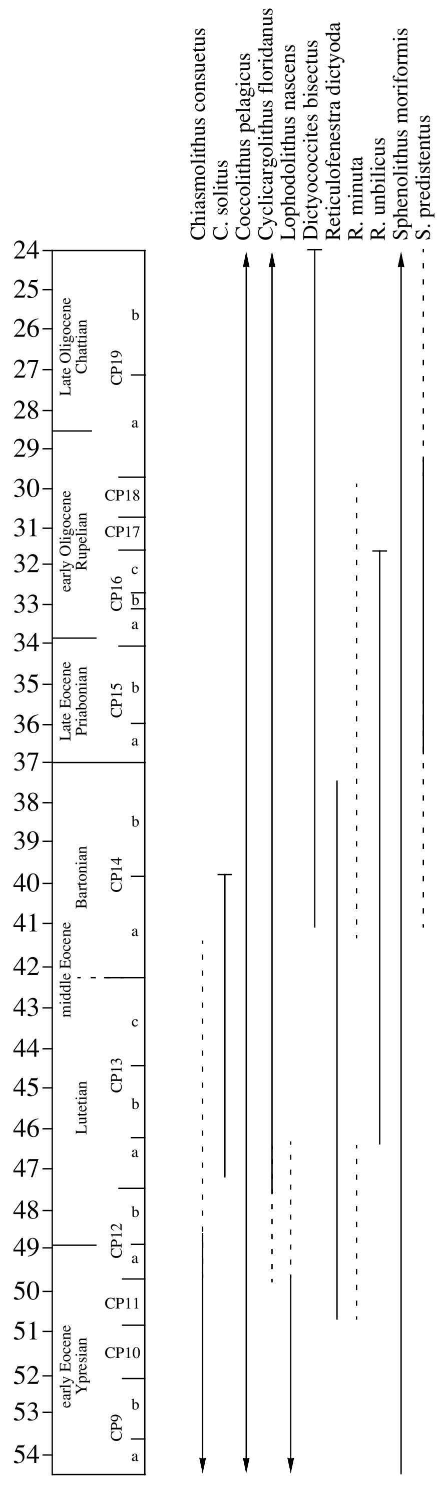

Fig. 3. Stratigraphic range of nannoplankton species found in terrigenous deposits in the northeastern part of the Karaginskiĭ Island (eastern Kamchatka).

The early–middle Eocene nannoplankton assemblage from the Supra-Khynkhlonai sequence substantially differs from the Late Eocene–early Oligocene assemblage found previously in the flysch complex

(Shcherbinina, 1997). The *in situ* nannoplankton from sedimentary melange, which was found in two samples from one and the same exposure, does not differ from nannoplankton of the flysch complex and corresponds to the Late Eocene–early Oligocene stratigraphic range.

The source of redeposited nannofossils and modes of their redeposition represent still an open issue. It is apparent that they are from some Late Cretaceous, mainly Santonian–Campanian and most likely poorly lithified sequence, in which a considerable part of nannoplankton was buried in pellets transportable after washing-out. The Vatyna and Irunei siliceous–volcanogenic sequences of Kamchatka and southern Koryak Upland, which are exposed in the vicinity of Karaginskiĭ Island, are mainly of the Santonian–Campanian age. After the arc–continent collision in the second half of the Eocene and Oligocene, the sequences were definitely subjected to intense erosion. However, nannoplankton has not been found in these strata. The provenance of sandstones and pebbles in flysch and melange, or origin of basaltic olistoliths in flysch and of tuff-basaltic lenses in melange, which are widespread in the southeastern part of the Karaginskiĭ Island, are unknown so far. It is likely that the Late Cretaceous redeposited nannoplankton occurring in the Eocene–Oligocene mudstones of the island was also related to a provenance undeterminable at present.

Hence, the case of Eocene–Oligocene deposits of Karaginskiĭ Island shows that redeposition of nannoplankton from older deposits to considerably younger ones is not only possible theoretically, but can be a significant event. If the nannoplankton content in rocks is low and number of collected samples is small, such a redeposition may cause large errors in dating the terrigenous sequences.

The fission-track dating of detrital zircon. The study carried out confirms in general the previous inference that ages of youngest zircon populations characterize, as a rule, stratigraphic ranges of their host sandstones, which accumulated in tectonically active zones (Brandon and Vance, 1992; Garver and Brandon, 1994; Solov'ev *et al.*, 2001; Shapiro *et al.*, 2001). In particular, the zircon ages of tephroids from the Supra-Khynkhlonai sequence (50.2 ± 3.2 Ma) are concordant with the sequence age established based on nannoplankton (the late Ypresian–early Lutetian). The age of tuffaceous sandstone from sedimentary melange (26.1 ± 1.5 Ma, Sample Ka22) is very close to the age of nannoplankton from matrix of adjacent pebbly mudstones (Sample Ha21, Bartonian–early Oligocene). The age of most zircon grains in sandstones from the flysch complex falls within the interval of 46–30 Ma and corresponds to the flysch age based on nannoplankton (Shcherbinina, 1997). On the other hand, the young zircon population in the melange complex is younger (26–18 Ma) than the complex age established on the basis of radiolarians (Chekhovich *et al.*, 1990). This is likely related to the fact that cherts in the melange complex are represented by olistoliths, which are older than the matrix.

The age similarity between young zircon populations and sandstones enclosing them implies that the zircon grains are genetically related with either the synchronous volcanism or the very fast exhumation and cooling of intrusions in the provenance. Some sandstones from sedimentary melange (e.g., Ka22) do contain substantial tuffaceous admixture, whereas most sandstones from the flysch complex of the Karaginskiĭ accretionary prism lack any tuffaceous material. More than half of them is made up of acid volcanics and granitoids and their derivatives (quartz and plagioclase) (Shapiro, 1984; Shapiro *et al.*, 2000). Hence, acid magmatic rocks of the provenance may be thought to be the sources of zircon grains. The inference validity can be tested, as the zircon fission-track ages are established not only for sandstones but also for gravelstones of the flysch complex, which are composed of acid volcanics, and for rhyolite and granite pebbles and boulders from pebbly mudstone of the flysch and sedimentary melange (Ledneva *et al.*, in press). It has been found that zircon grains in gravelstones are of the same age like those in sandstones. In granite pebbles, zircons are of Paleocene age, whereas rhyolite pebbles from the same rock are of the Maastrichtian age according to zircon dating. The age difference between granites and rhyolites is easily accounted for by necessity of a certain time-interval for granite exhumation. It can be expected therefore that combined washout of these rocks would give two zircon populations: one corresponding to the Cretaceous volcanism and the other one corresponding to the uplift of shallow intrusions at the surface. The integration of both populations into one with the mode of about 65 Ma is also a possibility. A zircon population with the last age approximately is truly established for some sandstones and gravelstones (Ka9, Ka11). Such zircon grains are less abundant than zircons subsynchronous to sedimentation.

Nonetheless, there are lenticular sandstone and gravelstone (Ka20) bodies in the sedimentary melange complex, in which the first zircon population is 70 Ma old. These are either olistoliths of Cretaceous sandstones or fragments of sandy layers similar in age to other sandstones but barren of zircon making up the young population.

The only fission-track age available for tuffs from the sedimentary melange complex (Ka18) corresponds to the beginning of the Campanian (82 Ma). The most probable explanation in this case is the tuff occurrence in the form of large and small olistoliths.

The data obtained confirm the earlier inference (Chekhovich *et al.*, 1990) that terrigenous rocks, which make up the bulk of the flysch and sedimentary melange complexes in the southeastern part of Karaginskiĭ Island, accumulated between the second half of the middle Eocene to the beginning of the Miocene. Sediments in these complexes become younger from northwest to southeast, i.e. from the continent to the past trench. The youngest fission-track ages were obtained for sandstones from the sedimentary melange complex.

Formation of the accretionary prism. The initial and terminal formation stage of the Karaginskiĭ accretionary prism did not left geological records. Based on the accretionary prism formation model, it may be suggested that imbricated thrust faults and folds in the flysch complex, like the chaotic complexes of sedimentary melange, are subsynchronous to sedimentation. In this case, the upper part of accretionary prism exposed on the island corresponds to the middle Eocene–early Oligocene subduction under the Kamchatka Peninsula. During that time, the Kinkil (or West Kamchatka–Koryak) volcanic belt formed on the isthmus of the Kamchatka Peninsula (Gladenkov *et al.*, 1997; *Explanatory Notes*…, 2000). At the end of the Eocene to the very beginning of the Oligocene, volcanism was likely active in the Govena arc. However, geological connection of the Karaginskiĭ prism with the Govena arc is doubtful. First, it is unlikely that the Govena arc extended to Karaginskiĭ Island since the Oligocene, because this interval in the island section is represented by terrigenous sediments. Second, the Karaginskiĭ prism is situated too close to the axial zone of the island, where the Govena arc extension is assumed to be. The present-day fore-arc structures more than a hundred kilometers wide are situated between the accretionary prism and the volcanic arc.

The youngest sandstones in the sedimentary melange complex enable a suggestion that the Karaginskiĭ accretionary prism was under formation up to the beginning of the Miocene, and developed even later in its underwater part.

CONCLUSION

(1) The studied mudstones of the northeastern part of Karaginskiĭ Island bear two nannoplankton assemblages of a low diversity and abundance. We consider the first assemblage ranging in age from the Ypresian–Lutetian (mudstones overlying volcanics of the Khynkhlonai Formation) to Bartonian–Oligocene (the sedimentary melange complex) as occurring *in situ* and characterizing age of sediments. The second assemblage of Late Cretaceous–Paleocene forms signifies rewashing of older sediments within the unknown provenance. The redeposited nannofossils are comparable with the *in situ* assemblage in abundance, diversity, and preservation state of species.

(2) According to the fission-track dating, the age of youngest detrital zircon population is close to sedimentation time of their host rocks. In the Supra-Khynkhlonai sequence, such a population corresponds in age to the Ypresian–Bartonian boundary that is consistent with the nannoplankton age of Sample F. In the flysch complex, the zircon populations range in age from 46 Ma (Bartonian) to 30 Ma (early Oligocene) that also corresponds to the sequence age based on nannoplankton (Shcherbinina, 1997). Data on the sedimentary melange complex are less compelling, though when it is possible to establish simultaneously the nannoplankton age of mudstones (Ha21, Bartonian–early Oligocene) and the age of young zircon population in adjacent tuffaceous sandstones (Ka22, 26.1 ± 1.5 Ma), the data do not contradict each other.

(3) In the northeastern part of Karaginskiĭ Island, terrigenous deposits become younger in the southeastward direction from the axial zone composed of Cretaceous and early Paleogene island-arc volcanics. Samples taken from the flysch complex are of the Bartonian–early Oligocene age, whereas the fission-track dating of detrital zircons suggests the Oligocene–early Miocene age of sedimentary melange complex. The situation when rocks become younger toward a trench and rock beds dip predominantly in the opposite direction is typical of accretionary prisms.

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