South Sakhalin tectonics and geodynamics: A model for the Cretaceous-Paleogene accretion of the East Asian continental margin

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Received 15 July 2005; revised 10 September 2005; accepted 1 October 2005; published 3 November 2005.

[1] This paper presents new data for the geological structure of South Sakhalin which is a connecting link between the structural features of the Sakhalin and Hokkaido islands which form the Late Mesozoic–Early Cenozoic Hokkaido-Sakhalin fold system. These data were used as a basis for deriving a new model of the tectonic evolution of the region, derived on the basis of terrain analysis. They illustrate the evolution of the East Asia continental margin during the larger part of the Cretaceous and during the Paleogene. The tectonic data available for this region recorded five epochs of structural readjustments, which separated the periods of the accretional growth of the continental margin. It is shown that the geodynamic evolution of the East Asian continental margin was controlled by the combination of convergence and transformation environments at the continent-ocean boundary and was complicated by the accretion of large intraoceanic volcanic rises and by the collision of ensimatic island arcs. *INDEX TERMS:* 1209 Geodesy and Gravity: Tectonic deformation; 1744 History of Geophysics: Tectonophysics; 3040 Marine Geology and Geophysics: Plate tectonics; *KEYWORDS:* East Asian Continental Margin, Geodynamics, Cretaceous-Paleogene accretion, tectonics.

Citation: [2] Zharov, A. E. (2005), South Sakhalin tectonics and geodynamics: A model for the Cretaceous-Paleogene accretion of the East Asian continental margin, *Russ. J. Earth. Sci., 7,* ES5002, doi:10.2205/2005ES000190.

Introduction

[3] In modern tectonic studies, the East Asian continental margin is treated as a collision-accretion feature, produced by the complex combination, in space and time, of accretion and collision processes, as well as of the transform fault movements of both individual terrains and their assemblages [*Bogdanov and Khain,* 2000; *Chekhovich,* 1993; *Golozubov,* 2004; *Isozaki,* 1996; *Khanchuk,* 1993; *Maruyama et al.,* 1997; *Natal'in,* 1991; *Parfenov et al.,* 1999, 2003; *Rozhdestvenskii,* 1993; *Sokolov,* 1992, 2003; *Sokolov et al.,* 1997; *Zonenshain et al.,* 1990].

[4] One of the key positions in the structure of the continental margin is occupied by the terrains of the accretion prisms of different ages (Figure 1), which recorded the time and mechanisms of the continental growth. The subduction, turbidite, and island-arc terranes, associated with them, produced the lateral series of structural features and allow one to reconstruct the old continent-ocean transition zones. In the modern structural pattern of the continental margin, these lateral rock series are seen to have been disturbed significantly by the late Mesozoic large size strike-slip movements along the margin, and by the Cenozoic destructive pull-apart and strike-slip movements which operated during the opening of the Sea of Japan and the Sea of Okhotsk basins.

[5] To a lesser extent these movements complicated the structural features of the Late Mesozoic-Cenozoic Hokkaido-Sakhalin fold system, represented by the structural features of the Sakhalin and Hokkaido islands, a connecting link between them being South Sakhalin. Over the whole length of this fold system, its eastern segments, including those of South Sakhalin, represent a complex zone of the connection of the oceanic structural features and those of the continental margin, providing good objects for reconstructing the evolution of the transitional zone.

[6] Presented in this paper are the new data available for the geology and tectonics of Southeast Sakhalin, which add to the knowledge of the structure and tectonic history of the Hokkaido-Sakhalin accretion-type fold system and offer new tectonic reconstructions for the East Asia continental margin for Cretaceous-Paleogene time.

ISSN: 1681–1208 (online)

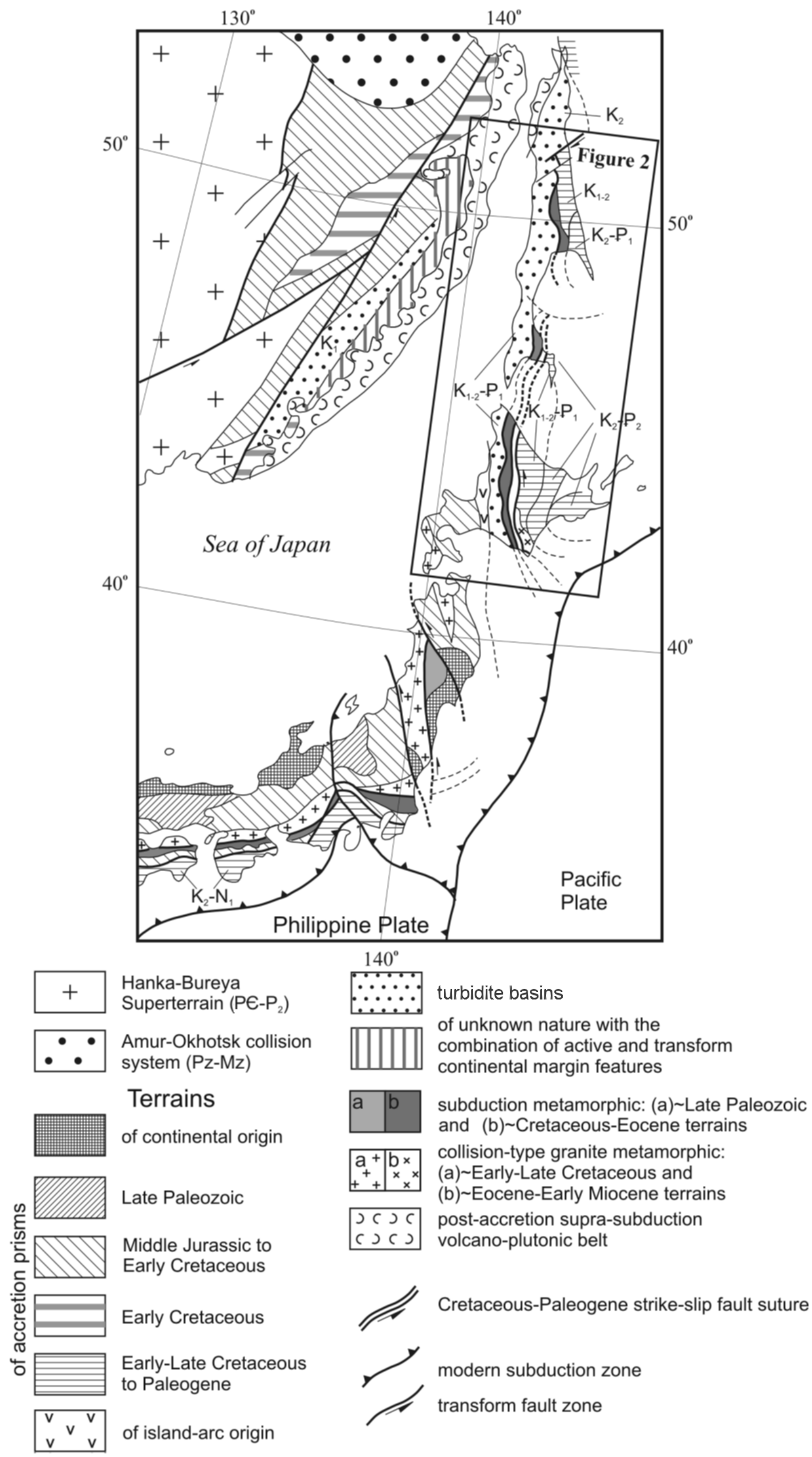

Figure 1. The position of the Hokkaido-Sakhalin fold system in the accretion-type structural features of the East Asia continental margin (compiled using the maps of *Natal'in* [1991], *Khanchuk* [1993], and *Isozaki* [1996]).

Tectonic Mapping of the Hokkaido-Sakhalin Fold System

[7] Most of the geologists, engaged in the study of this region, share the view of the well expressed transverse tectonic zoning of this fold system traceable from the southern areas of Hokkaido to the Shmidt Peninsula in the north of Sakhalin.

[8] **Tectonic mapping of Hokkaido.** This area is known for the most complete presence of its tectonic zones and was most thoroughly studied [*Dobretsov et al.,* 1994; *Isozaki,* 1996; *Kiminami et al.,* 1986, 1992; *Niida and Kito,* 1986, to name but a few]. Eight terrains differing in their structure, genesis, formation time, tectonic positions, and genesis, which reflect the successive growth of the continental margin from the Middle Jurassic to the Neogene, have been mapped there.

[9] The southwestern area of Hokkaido is occupied by the *Oshima Terrain* [*Kiminami et al.,* 1992] which is often compared with the Middle Jurassic (Berriasian) terrains of the accretionary prisms of the Central Sikhote-Alin area and with those of the Honshu Island (Samarka, North Kitakimi, and Mino-Tamba) [*Golozubov,* 2004; *Isozaki,* 1996; *Maruyama,* 1997].

[10] The more eastern area is occupied by the *Rebun-Kobato* linear island-arc terrain, which is often included in the Oshima Terrain [*Kiminami et al.,* 1986; *Niida and Kito,* 1986]. This terrain is dominated by Tholeiite andesite and basalt, as well as by volcanic sedimentary rocks, dated Berriasian-Barremian in the Kobato Mountains in the west of Hokkaido [*Nagata et al.,* 1986] and Valanginian-Barremian in the Rebun Island [*Ikeda and Komatsu,* 1986].

[11] The axial part of Hokkaido is usually referred to as the Sorachi-Iezo tectonic belt which shows the complex tectonic relations between the pre-Barremian oceanic volcanogenic siliceous rocks (Sorachi Group), the Hauterivian (?) – Barremian – Paleocene flysch, turbidite, and shallow-sea sediments of the Iezo Supergroup, their metamorphic analogs, and numerous serpentinite melange with ophiolite and high-pressure rock slices (Kamuikotan metamorphic belt). In this paper they are treated as independent tectonic units which satisfy the definition of a terrain [*Parfenov et al.,* 1999; *Sokolov,* 2003], namely, the Iezo Barremian-Paleocene turbidite terrain, the Sorachi (pre-Barremian oceanic plateau), and the Kamuikotan Cretaceous-Early Eocene subduction-type metamorphic terrain.

[12] The *Idonnappu suture zone (belt)* is the eastern restriction of the Central Hokkaido terrains. It consists of steeply standing slabs of turbidite, olistostrome, and melange, which grow younger eastward from Barremian-Albian to Campanian-Danian, and also of large oceanic basalt slabs of different origin [*Kiyokawa,* 1992; *Ueda et al.,* 1994]. The suture zone shows shear deformations, the earliest of them being ranked as left lateral with eastern vergence [*Kiyokawa,* 1992]. The Idonnappu zone is usually interpreted as a Cretaceous accretion, yet, the wide development of shear zones of different time periods, as well as the structured pattern of the melange and the linear pattern of the entire structural feature suggest its shear origin, the presence of large ophiolite allochthons allowing one to rank it as a suture-type structural feature.

[13] The *Hidaka Terrain* corresponds to the Late Cretaceous-Early Eocene accretion prism, composed of turbidite and east-vergence terrigenous melange zones [*Kiminami et al.,* 1992]. The orientation of the paleocurrents and the composition of the turbidite suggest that the clastic material had been derived from the side of the continental margin. The positions and origin of the eastern boundaries of this terrain are still a matter discussion.

[14] The *Tokoro Terrain* [*Kanamatsu et al.,* 1992; *Kiminami et al.,* 1992; *Sakakibara et al.,* 1986, 1993; *Tajika,* 1988] is composed of Campanian-Early Eocene hemipelagic rocks and tuff turbidite (Yubetsu Group), of Upper Cretaceous and Lower Paleocene conglomerates and turbidites (Saroma Group), and of Middle-Late Jurassic to Cenomanian siliceous volcanic rocks of oceanic islands (Nikoro Group). The Saroma and Yubetsu Groups are known to be treated as the rocks of the fore-arc trough or as those of the accretion prism, which had accumulated under the conditions of the Mezopacific oceanic plate subduction under the ensimatic island arc [*Kanamatsu et al.,* 1992; *Kiminami et al.,* 1986, 1992; *Kimura,* 1996]. In the southern part of the terrain the accretion rocks are overlain with an angular unconformity by a Middle-Late Eocene neoautochthon and were crowded, along with latter, during the Early-Middle Miocene in the course of the Nemuro island arc collision [*Bazhenov and Burtman,* 1994; *Kimura,* 1996].

[15] The *Nemuro Terrain* is exposed in the southeast of the Hokkaido Island and extends as far as the islands of the Minor Kuril Arc [*Kiminami,* 1983; *Kiminami et al.,* 1992; *Melankholina,* 1988]. The eastern part of the terrain is composed of Campanian-Late Eocene tuff turbidite and flysch with dolerite sills and island-arc volcanic flows. The western part of the terrain, contacting the Tokoro Terrain, is composed of Campanian-Middle Miocene flysch, turbidite, and conglomerates, and does not contain any effusive rocks. The compositions of the turbidites and volcanic rocks suggest their accumulation in the environment of an ensimatic volcanic arc. The results of the paleomagnetic studies carried out in the eastern part of the terrain [*Bazhenov and Burtman,* 1994] suggest the arc was formed in a low-latitude region and that its accretion to the continental margin took place only in the Early-Middle Eocene.

[16] The main specific feature of the subdivision of Hokkaido into tectonic regions was the fact that the western and central terrains, including the Hidaka accretion terrain, were ranked as the structural features produced by the subduction of the Paleo-Mezopacific oceanic plates under the continental margin. The origin of the eastern terrains is explained by the subduction of opposite polarity. Most of the authors associate their origin with the subduction of the oceanic crust under the inferred Sea of Okhotsk microcontinent [*Jolivet,* 1986; *Kiminami,* 1983; *Kiminami et al.,* 1986, 1992, to name but a few]. However, some geologists suggest that the Tokoro and Nemuro terrains were formed in the environment of an ensimatic island arc [*Bazhenov and Burtman,* 1994; *Bazhenov et al.,* 2001; *Kimura,* 1996].

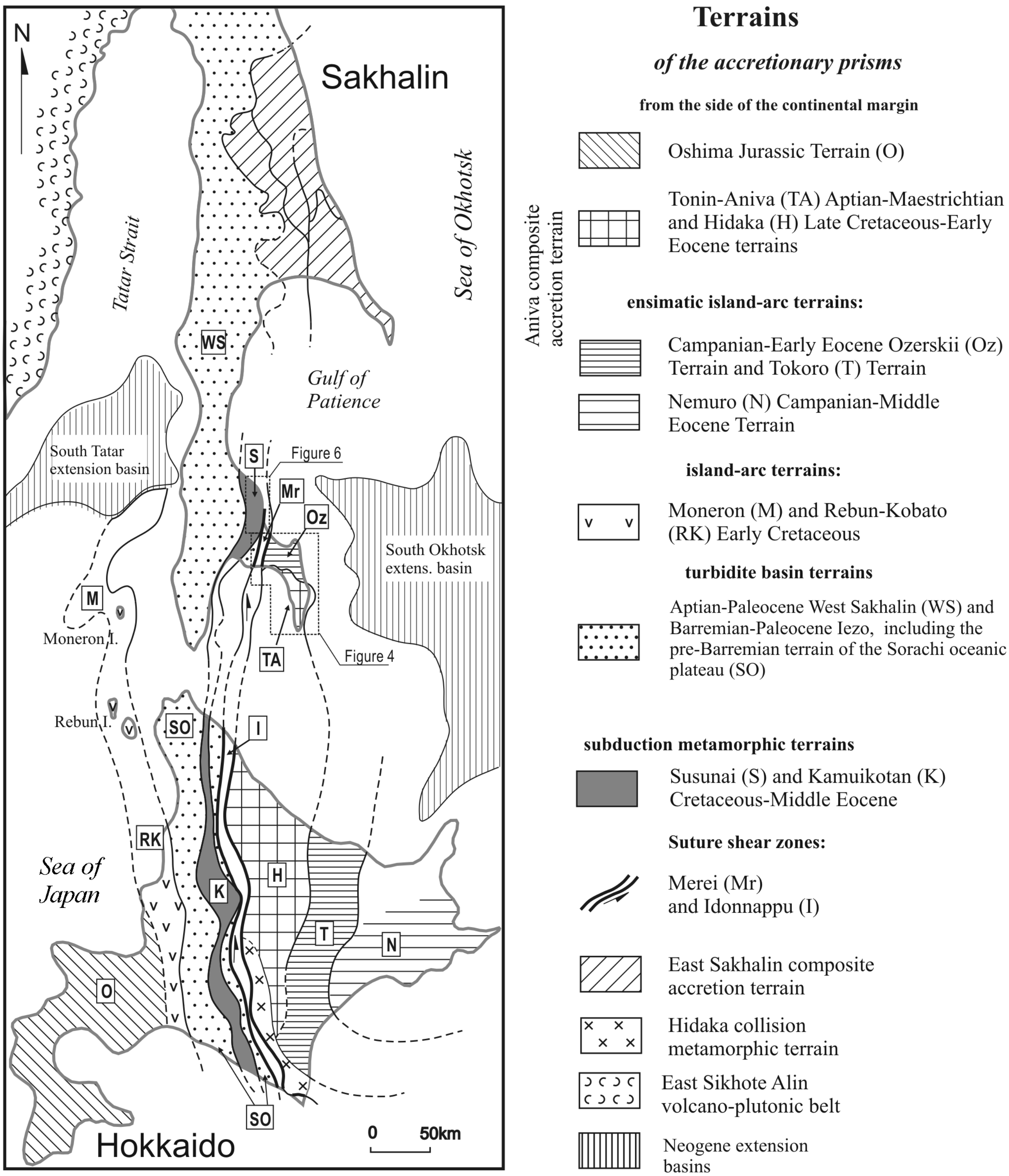

Figure 2. Tectonic units of the Hokkaido-Sakhalin fold system.

[17] The tectonic units of Central Hokkaido extend conventionally into South Sakhalin, whereas the eastern terrains are believed to be unique for Hokkaido, or are correlated with the eastern segment of Central Sakhalin.

[18] **The tectonic mapping of South Sakhalin** was based on the analysis of the known data and on the data reported in this paper. The following terrains were identified, from west to east, in South Sakhalin and in the adja-

cent sea areas: the Moneron Neocomian island-arc terrain, the West Sakhalin Aptian-Paleocene turbidite terrain, the Susunai Cretaceous-Early Eocene metamorphic subduction terrain, and the polygenic terrains of the accretion prism, namely, the Tonin-Aniva Aptian-Maestrichtian terrain and the Ozerskii Campanian-Early Eocene terrain (Figures 2 and 3). The terrains of the accretion prisms are separated from the more western ones by the Merei left-lateral suture zone.

[19] The *Moneron Terrain* is interpreted as the northern continuation of the Rebun-Kobato Neocomian island-arc terrain. Its structure was studied in the Moneron Hole [*Piskunov and Khvedchuk,* 1976], which penetrated the three-kilometer sequence of island-arc tholeitic and calc-alkaline volcanic rocks ranging from intermediate to basic compositions, and including thin members of volcanogenic sedimentary rocks. This volcanic rock complex is layered tectonically into several members with a Cretaceous-Paleocene age range, which are combined irregularly in different parts of the rock sequence, and grouping generally into the following four age intervals: 141, 118, 103–86, 77–59 Ma.

[20] The Aptian-Maestrichtian turbidite and the Paleocene coastal-marine and paralic coal-bearing rocks of the *West Sakhalin Terrain*, totaling more than 6 km in thickness, correlate confidently with the rocks of the Iezo Terrain. The stratigraphic and tectonic specifics of this terrain were discussed in detail in literature [*Governmental Geological Map,* 2001; *Melankholina,* 1973, 1988; *Zyabrev,* 1984, 1987, 1992]. The eastern flank of this terrain is poorly metamorphozed at the contact with the Susunai Terrain of metamorphic rocks. In the SE direction the distal turbidite facies are crushed in the Merei shear zone and transformed to mylonite. The Upper Cretaceous-Paleocene terrigenous rocks of the West Sakhalin terrain are overlain by Eocene-Oligocene deposits, with a hidden stratigraphic unconformity in the west, and with a distinct angular unconformity in the east [*Governmental Geological Map,* 2001; *Zharov et al.,* 2004]. In the north-western part of their outcrop, these rocks are cut by a belt of Paleocene biotite rhyolite dikes of subduction-collision origin. It follows that the West Sakhalin terrain is a fragment of the Aptian-Paleocene continental margin trough, bounded in its flanks by deep long-lived fold zones. The closure and inversion of this trough is clearly seen at the Cretaceous-Paleogene boundary.

[21] The *Susunai Terrain* is interpreted as the northern continuation of the Kamuikotan metamorphic terrain, being comparable with the latter both in the age of the metamorphism and in the tectonic position and internal structure. The metamorphic rocks of these terrains are more often interpreted as the deep levels of the Cretaceous accretion rocks, or as the surfaces of the Cretaceous subduction zones [*Dobretsov et al.,* 1994; *Kimura et al.,* 1992b; *Komatsu et al.,* 1992; *Sakakibara and Ota,* 1994]. As follows from another viewpoint [*Richter,* 1986], the metamorphic structure was formed as a result of the Maestrichtian-Paleocene metamorphism of the turbidites of the West Sakhalin terrain and its basement in the course of the obduction of the perioceanic structural features over the continental margin.

[22] A radically new interpretation of the geotectonic structure is offered in this paper for the *Tonin-Aniva* Peninsula (Figure 3). In contrast to the previous views, only the central and southern parts of the peninsula are comparable with the Khidaka Terrain and can be interpreted as the Tonin-Aniva Terrain of the continental-margin accretionary prism.

[23] The northern and northeastern parts of the peninsula are treated as the *Ozerskii Terrain* with its widely developed rocks of the Permian-Cenomanian oceanic crust of the Paleo- and Meso-Pacific Ocean and the fragments of the Campanian-Early Eocene ensimatic island-arc system and of the Late Cretaceous-Paleocene sedimentary cover of the epi-oceanic marginal sea. The Ozerskii Terrain was correlated with the Tokoro Terrain and was found to be the accretionary prism of the ensimatic island arc [*Bazhenov et al.,* 2001, 2002; *Zharov,* 2003].

[24] The Tonin-Aniva and Ozerskii terranes are separated by the Vavai melange zone extending from the western part of the peninsula to its southeastern end. All over the boundary melange area, the structural features of the Ozerskii Terrain overlap the Tonin-Aniva Terrain, proving the allochthonous origin of the former.

[25] The Tonin-Aniva and Ozerskii terrains form the Aniva composite terrain [*Zharov,* 2003, 2004], the origin of which was dated Middle-Late Eocene, proceeding from the age of the Vavai melange and of the suturing collision granitoids.

[26] The *Merei suture-type shear zone* is the recently mapped tectonic element of the region, which is comparable in many respects with the Idonnappu suture zone. These structural features are controlled by extensive shear sutures, separate the Cretaceous-Early Paleogene accretion and turbidite terranes, and are accompanied, along the western flank, by the left-lateral en-echelon outcrops of high-pressure rocks and melanged ophiolites of the Kamuikotan and Susunai terrains.

[27] The comparison of the tectonic elements of South and Central Sakhalin shows that only one West-Sakhalin Terrain can be traced reliably in the northern direction. Its Late Cretaceous distal turbidites are replaced at the latitude of the northern part of the Gulf of Patience by proximal facies and grade further northward to coastal marine and paralic coal-bearing rocks [*Melankholina,* 1973, 1988; *Zyabrev,* 1987, 1992]. This suggests the discordant orientation of the modern boundaries of the terrain relative to the Late Cretaceous structure-facies zoning of the continental margin trough. The metamorphic and accretion-related rocks are widespread also in Central Sakhalin, east of the West Sakhalin Terrain. Identified among these rocks were the equivalents of the Tonin-Aniva Terrain and, less confidently, some tectonostratigraphic units of the Ozerskii and Susunai terrains [*Khanchuk,* 1993; *Richter,* 1986; *Rozhdestvenskii,* 1993, to name but a few]. However, the differences in the structure and tectonic positions of individual rock complexes, as well as the different Paleogene evolution of South and Central Sakhalin preclude any correct correlations. In this paper the metamorphic, accretion-type, and sea-margin rocks of Central Sakhalin are classified as the *East-Sakhalin composite terrain* of accretion-collision origin, which developed from the end of the Early Cretaceous to the Eocene, inclusive.

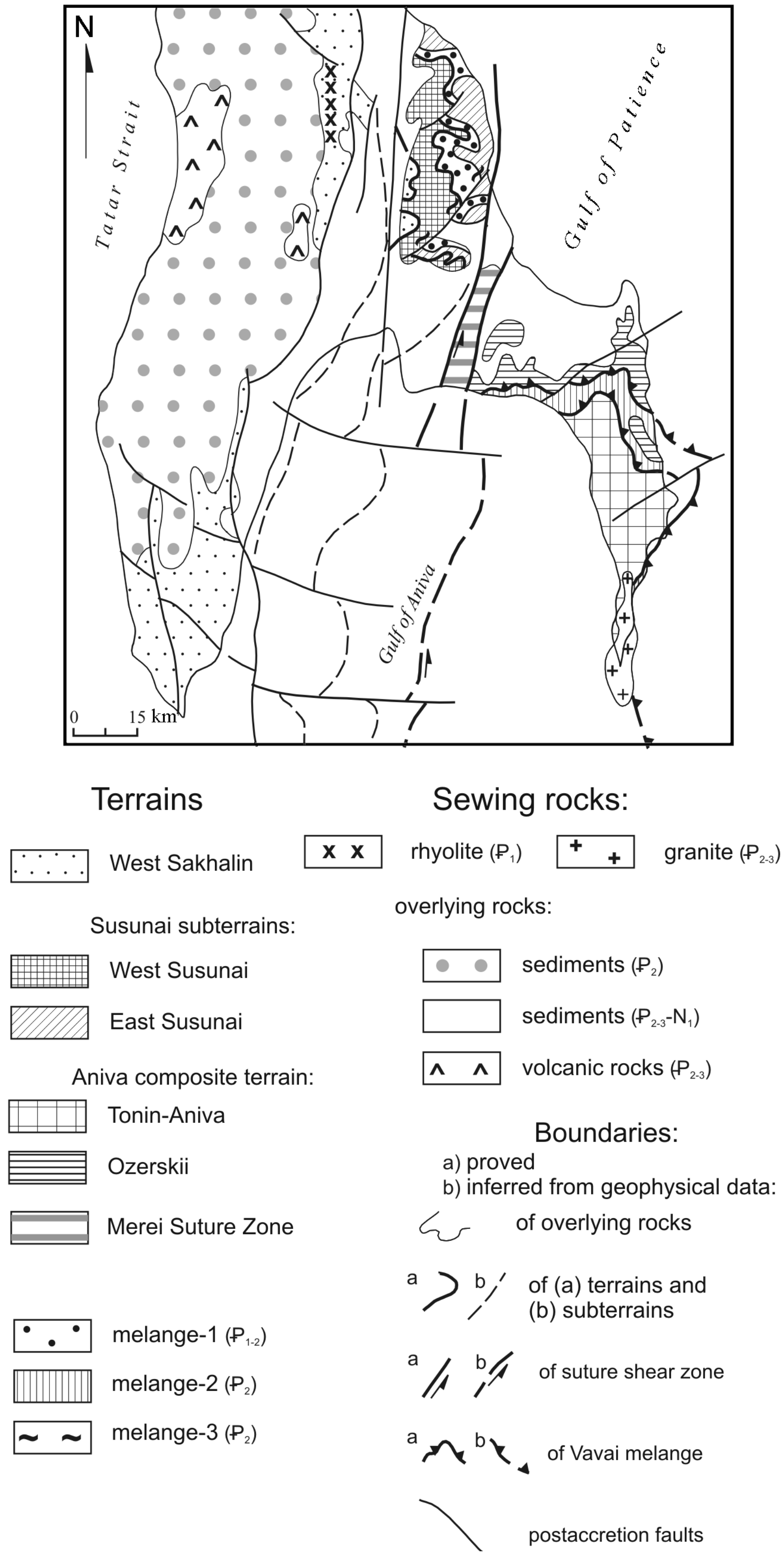

Figure 3. Map of the South Sakhalin terrains.

[28] Discussed below are the main lithologic and structural features of the Tonin-Aniva, Ozerskii, and Susunai terrains and of the Merei suture zone, which illustrate the tectonic types and collision structures of South Sakhalin, as well as, the models for the formation and evolution of the Cretaceous-Paleogene East Asian continental margin.

South Sakhalin Accretion and Collision Structural Features

[29] **Accretion-type structural** features are developed in the Tonin-Aniva Peninsula, where they occur as the Tonin-Aniva, Ozerskii, and Susunai accretion terrains, and also in the Susunai Range, where the Susunai metamorphic subduction terrain is exposed.

[30] **The Tonin-Aniva Terrain** is composed of Middle and Late Cretaceous turbidite and melange-olistostrome rocks with the distinct progradation of the clastic rock series in the eastern direction (Figures 4 and 5). Its main rocks are the Aptian-Cenomanian turbidites and melanged olistostromes, combined into the Utesna Sequence and the coherent sequences of the Upper Cretaceous flyschoid turbidite, ranked as the Evstafiev Formation. The melanged olistostromes include large allochthonous slabs of Jurassic-Lower Cretaceous volcanogenic siliceous terrigenous rocks, identified as a hard rock series. The accretion rock complexes of this terrain are intruded in the southern part of the peninsula by Middle Eocene-Early Oligocene S-type granites and are overlain, in graben-shape depressions, by Lower-Middle Miocene coal-bearing subcontinental deposits with thick weathering crust at the base.

[31] The common features of the hard-rock series are: (1) the siliceous volcanic composition of its lower part and the volcanogenic terrigenous composition of its upper part; (2) the combination of Tholeiite and subalkaline basalts in its lower and middle parts and the predominance of alkaline basalt, up to picritic ones, in the upper part; (3) the association of siliceous and carbonate rocks with the formation of jasper-limestone flyschoid alternation; (4) the predominance of quartz-jasper and low-quartz feldspar with albitophyre-andesite-basalt clastic material. The radiolaria finds restrict the age of this rock sequence to the Middle Jurassic (Bathonian)–Early Cretaceous interval.

[32] The *Utesna rock sequence* has a chaotic structure because of widely developed olistostromes and imbricate dislocation zones grading to melange. The olistoliths are dominated by hard rocks. The sandstones vary from arkose to graywacke ones, which suggests different sources the clastic material, both oceanic and continental. The high-quartz rocks are typical of the turbidites, and occur as olistoliths in the olistostromes.

[33] The thickness of the slabs is not lower than 1 km, the total tectonic thickness of the accretion rock complex may be equal to or higher than 4–5 km, which is a typical value for accretion prisms. The radiolaria assemblages from the olistoliths and from the matrix date this rock complex Aptian-Cenomanian.

[34] The structural pattern of the Utesna rock sequence allows one to rank it as the Middle Cretaceous accretion rock complex that had accumulated during the accretion of the oceanic volcanic plateau or of some submarine volcanic mountains to the Asian continental margin. The presence of high-quartz and quartz feldspar sandstones in the turbidites, and of exotic Permian limestones in the olistostromes, suggests that this accretionary prism originated south of its modern position, in the vicinity of the outcrops of the old accretionary rock complexes or of large continental blocks of the Asian margin (for instance, the Northern and Southern Kitakami terrains of the Northern Honshu Island), unknown in the vicinity of South Sakhalin.

[35] The *Evstafiev Formation* is composed of siltstone, sandstone, tuffstone, and the members of their flyschoid interbedding. The lower part of this formation is dominated by siliceous aleuropelite with siliceous sandstone interbeds. Up the sequence the silica content declines, the rocks grow more tuffaceous, the rock sequence being dominated by flyschoid members with gradational bedding. The sandstones of this formation are distinguished by a low quartz content, being dominated by acid-intermediate igneous rock particles. The maximum thickness of this formation is 2600 m. The scarce Campanian-Maestrichtian radiolaria and the remnants of Late Cretaceous inocerams restrict the age of these rocks to Late Cretaceous. The specific structure of this formation suggests that its rocks had accumulated in the axial part of a deep-sea trench or in the lower part of its internal slope under the conditions of the eastward progradation of some forearc trough.

[36] This terrain has a scale-overthrust, intricately folded divergent structure which was shaped during several tectonic stages and is distinguished by the submeridional strike of its melanges and en-echelon overthrusts (see I–I in Figure 5).

[37] The overthrusts and melanges of eastern vergence controlled the growth of the tectonic thickness of the Utesna rock sequence and its thrusting over the Late Cretaceous turbidites in the eastern direction. The overthrusts and terrigenous melanges of the western vergence are developed mainly in the field of the Evstafiev rock development. In the central part of the peninsula they controlled the overthrusting of the Utesna and Scalnaya rock sequences in the western direction. The melange inclusions are composed mainly of siliceous rocks, the matrix being similar to the Late Cretaceous turbidites.

[38] **The Ozerskii Terrain** is characterized by the wide development of the rock complexes of the Permian-Early Cretaceous oceanic crust of the Paleo- and Meso-Pacific ocean, which are combined tectonically with the fragments of the Campanian-Early Eocene ensimatic island-arc system and with those of the Late Cretaceous-Paleocene sedimentary cover of the epioceanic marginal sea. The island-arc rocks are developed mainly in the eastern part of the terrain with its NNW structural trends, the marginal sea rocks developed in the west being distinguished by the structural features of the latitudinal strike. These specific features were used as a basis for the mapping of the eastern (Chaika) and western (Tunaicha) subterrains. The tectonic structure of this terrain in the northern part of the peninsula is masked

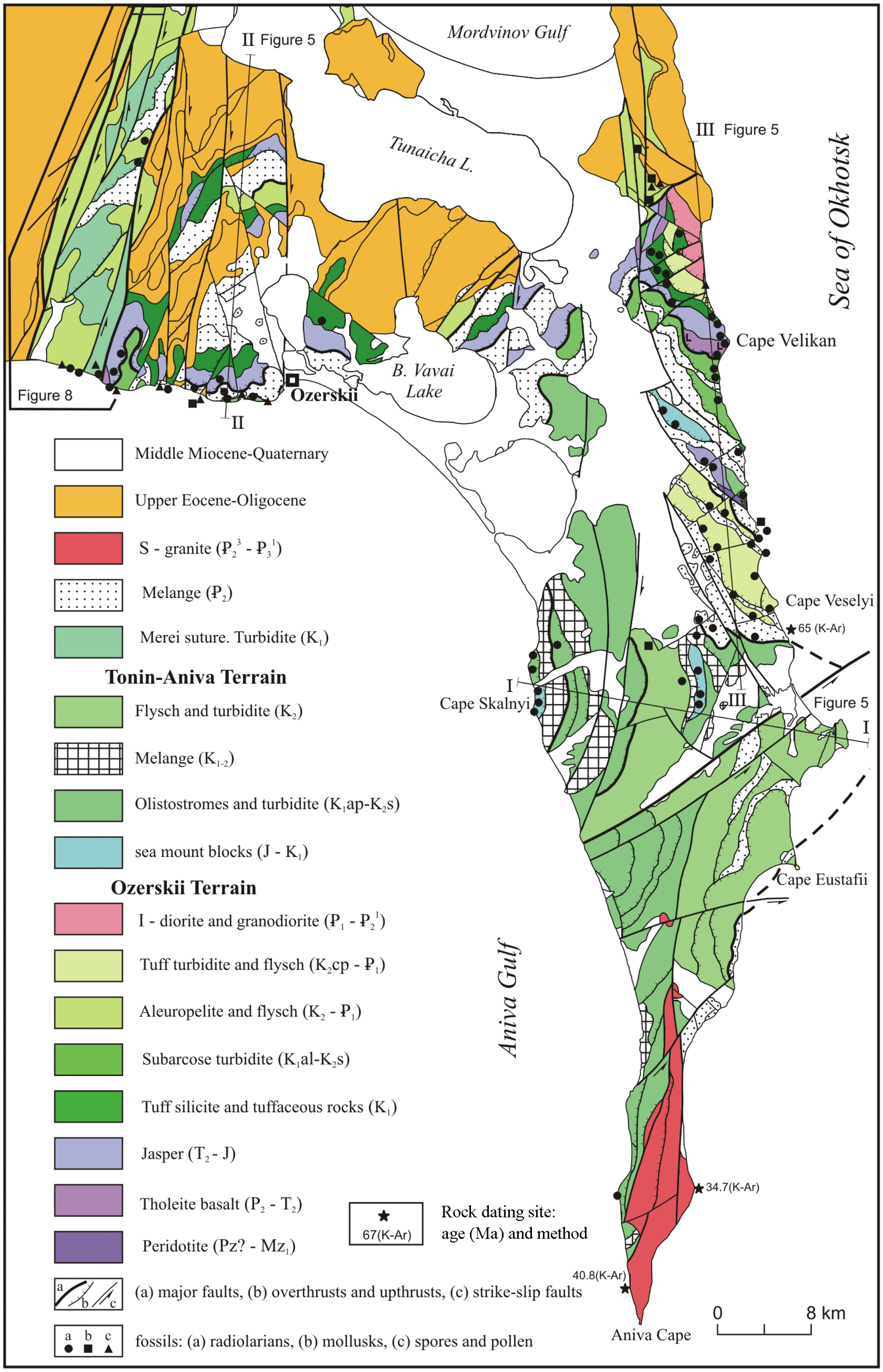

Figure 4. Schematic geological map of the Tonin-Aniva Peninsula.

by the Late Eocene-Early Miocene neoautochthon.

[39] The Ozerskii Terrain consists of six lithostratigraphic units embracing the time interval from the Late Permian to Paleocene, inclusive. It shows a persistent tectonic and stratigraphic sequence which was studied most thoroughly in the Chaika subterrain.

[40] The bases of both subterrains include the tectonic slabs of the Albian-Cenomanian subarcose turbidite of the Gorbusha Sequence, which is underlain by the Vavai melange and is comparable in terms of its sandstone composition with the coherent turbidite of the Tonin-Aniva Terrain (Figures 4 and 5). This rock sequence is composed of quartz-feldspar sandstone, siltstone, and members of their flyschoid intercalation with gradation stratification. The bottom of this rock sequence is dominated by sandstone with acid effusive and granite clastics (up to 35–40%), whereas the top is composed of sandstone with albitite-jasper and basaltic andesite composition of the fragments. The thickness of these rocks is higher than 1000 m.

[41] The subarkose turbidites are overlain by a tectonic layered oceanic rock complex composed of the Velikan, Yunona, and Kedrov rock sequences. The contacts between these rock sequences are stratigraphically concordant but deformed by faults.

[42] The *Velikan rock sequence* is complicated by metabasats, clastic hyalobasalt lavas and hyaloclastites with subordinate jasper layers at the top. In terms of their mineral and chemical compositions, the basalts are ranked as oceanic Tholeiite, and in terms of their minor elements, they resemble MOR basalts. The Late Permian-Middle Triassic (Anisian) age of these rocks was proved by the rare radiolarias found in the jasper interbeds. This rock sequence is more than 300 m thick.

[43] The *Yunona rock sequence* (Middle Triassic-Jurassic) is the main marker of the Ozerskii Terrain because of its permanent jasper composition. The jasper includes phthanite layers, radiolarites, and siliceous argillites in the upper part of this rock sequence, and occasional basalt lenses. The rock sequence of the Tunaicha subterrain (Yunona Mt. area) is ranked as a stratotypical one and has been subdivided into stages using 15 radiolarian complexes [*Bragin,* 1986, 1991]. The Triassic rocks have a thickness of 80–100 m. The Jurassic rocks are 170–200 m thick, the total thickness of this rock sequence being 300 m.

[44] The *Kedrov rock sequence* is composed of tuffaceous siliceous rocks in its lower part, which grade facially from tuffaceous silicite to cherty tuffite rocks with tuff members and subalkaline basalt lenses. The upper tuffaceous-terrigenous part of this rock sequence varies from site to site in the amounts of tuff, tuffaceous siliceous and terrigenous rocks and in the number of their flyschoid interbedding. The petrochemical features of the basalts suggest their oceanic, intraplate origin.

[45] The Upper Tithonian-Albian age of this rock sequence was determined using abundant radiolaria and more rare rudists and trigoniids. The thickness of this rock sequence varies from 800–850 m in the Chaika subterrain to 500 m in the Tunaicha one.

[46] The upper structural level in the Chaika Subterrain is occupied by the Chaika rock sequence which rests on the oceanic rock complex in argillaceous melange zones and is composed of tuffaceous turbidite and flysch of tuffaceous origin. Large rootless slabs of these rocks have been found in the Vavai melange. The rocks of this sequence show the vertical upward replacement of the hemipelagic deposits by thick-bedded terrigenous, very thick-bedded tuffaceous terrigenous rock members, with the markers of varicolored tuffaceous siltstone and tuffite, and tuffaceous-silty-sandy flyschoid members with olistostrome horizons in the top of the rock sequence. The olistoliths include both subalkaline and tholeitic oceanic basalts, and island-arc calcalkaline and Tholeiite basalts. This rock sequence varies from 850 m to 1100 m in thickness. Its psammites are low in quartz, contain some basaltic andesite clastic material, and correspond in their composition to ensimatic island arc sandstones. The rocks of this terrain contain abundant Campanian-Maestrichtian and Paleocene radiolarians and Early Paleogene spores and pollen.

[47] The *Late Cretaceous–Early Paleogene rocks* of the Tunaicha Subterrain have an aleuropelite and argillite composition and include rare flyschoid horizons and interlayers of tuffaceous sandstone in the eastern part of this subterrain. In some sections aleuropelite replaces the siliceous terrigenous rocks of the Kedrov rock sequence. These rocks are less than 650 m in thickness. This rock sequence was dated using scarce Late Cretaceous inocerams, Campanian-Maestrichtian radiolarians, and poor Early Paleogene spore and pollen spectra.

[48] The upper structural levels of the Chaika Subterrain are occupied by a zonal quartz diorite-granodiorite intrusion (Okhotsk Massif) and by the accompanying dike series of contrasting composition ranging from microdiorite to rhyolite and alkaline microgranite. These rocks have been classified as the Okhotsk diorite-granodiorite complex consisting of two phases. Ranked as the early phase are the rocks of this massif and the holocrystalline microdiorite and granodiorite porphyry dikes cutting only the Chaika rocks. They are composed of island-arc diorite and granodiorite, and their petrochemistry proves their ensimatic origin. The rocks of the second phase are acid and alkaline dikes cutting the rocks of the Okhotsk Massif, the dikes of the first phase, and the deposits of the Yunona, Kedrov, and Chaika rock sequences. The rocks of the second phase are fairly similar to those of the granite-syenite suite and correspond to the intermediate, subduction-collision, granitoids that had been formed during the collision of the island arcs with the continental structural features. The results of the K-Ar and zircon track dating restrict the ages of the early rocks to Paleocene-Early Eocene (63.5–48.5 Ma) and those of the second phase to Middle Eocene (45 Ma). The secondary alterations of the rocks were dated Miocene-Oligocene (37.4–22.3 Ma) and are believed to have been associated with the rise of the granites to the ground surface.

[49] The Ozerskii Terrain has a more ordered structural pattern, compared to the Tonin-Aniva one, and is distinguished by its differently oriented structural elements and by the deformation style of the Chaika and Tunaicha subterrains.

[50] The *Chaika Subterrain* has a steeply standing slab structure of the western and southwestern vergence and con-

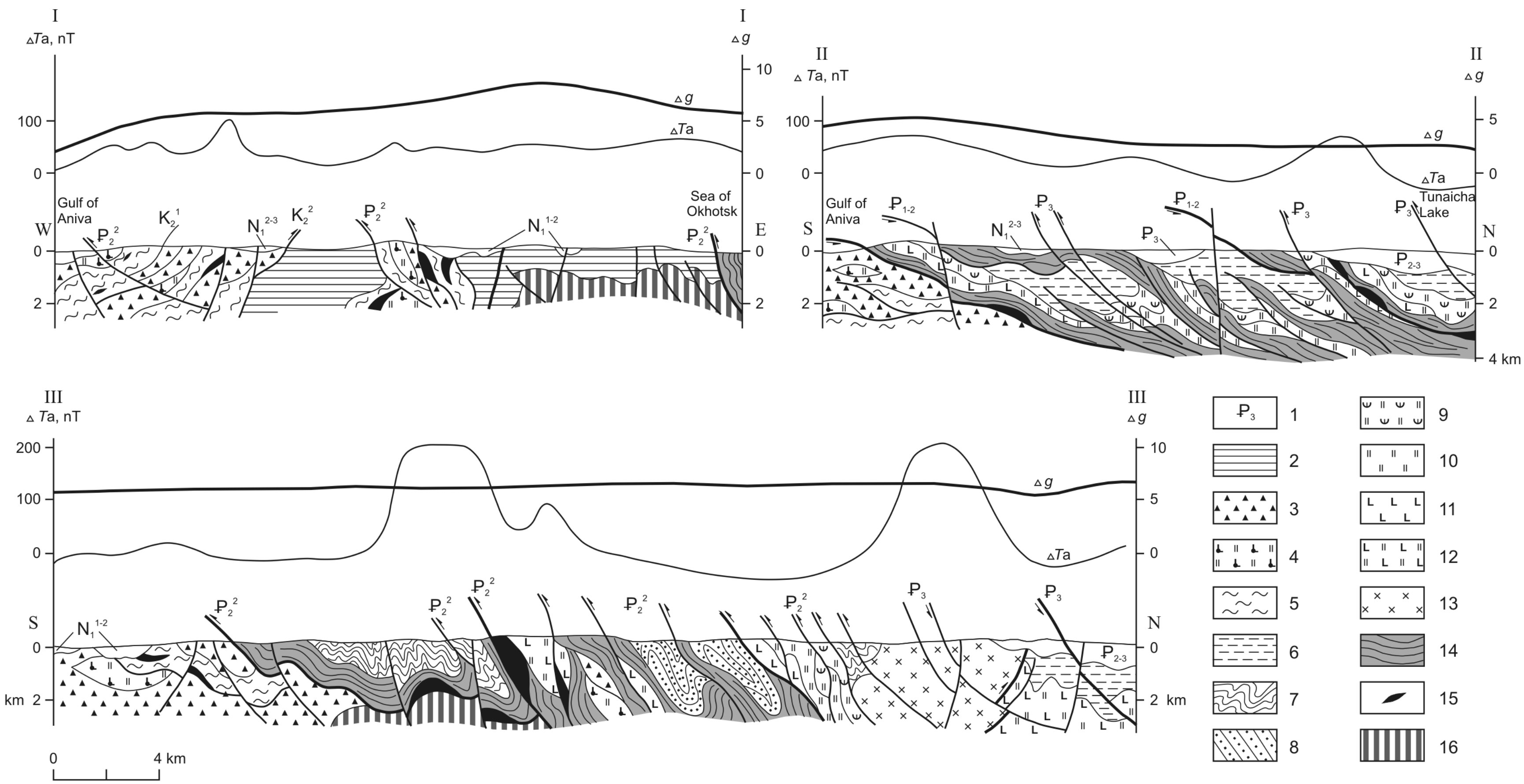

Figure 5. Geological profiles across the Tonin-Aniva Peninsula. (1) neoautochthon, (2–5) Tonin-Aniva Terrain: (2) Evstafiev Formation (K_2), (3) Utesna rock sequence (K_{1-2}), (4) hard rock sequence (J-K_1), (5) Ostrovskii melange (K_{1-2}), (6–14) Ozerskii Terrain: (6) flysch and aleuropelite (K_2-₽$_1$), (7) Chaika rock sequence (K_2-₽$_1$), (8) Gorbusha rock sequence (K_{1-2}), (9) Kedrov rock sequence (J_3-K_1), (10) Yunona rock sequence (T_2-J), (11) Velikan rock sequence (P_2-T_2), (12) jasper and basalt, (13) Okhotsk granodiorite (₽$_{1-2}$), (14) Vavai melange (₽$_2^2$); (15) serpentinite; (16) melanocratic basement.

sists of three slab packages, namely, the Gorbusha, Velikan, and Okhotsk (see III–III in Figure 5).

[51] The lower, Gorbusha slab package is composed of the subarcose turbidite of the Gorbusha rock sequence, overturned to the southwest. The Velikan slab package rests on the Gorbusha slab package along a melange zone with the slabs of the Gorbusha sandstone and volcanic siliceous rocks, comparable with the hard rocks of the Tonin-Aniva Terrain in terms of their age, lithology, and geochemistry. At the base of the Velikan package, the Velikan basalts are deformed to large, SSE-overturned folds and are broken tectonically into slabs, as thick as 100–150 m. The jasperoid rocks of the Yunona rock sequence compose the bulk of this tectonic package and are deformed to large recumbent and diving folds with the spans of their limbs measuring up to a few hundred meters. The northern limbs show the growth of the jasperoid rock sequence at the expense of the tuffaceous siliceous of the Kedrov rock sequence. The total thickness of this oceanic rock complex is not larger than 1.2 km, the total thickness of this slab package being larger than 3.5 km.

[52] The Velikan slab package is overlain, via the zone of scaly schist and tectonite, by the Okhotsk slab package composed of the Chaika tuffaceous terrigenous rock sequence cut by the granitoids of the Okhotsk Complex. The dikes of the first phase of the Okhotsk Complex were emplaced before the formation of the plate structure of the subterrain, whereas the rocks of the second stage showed their subduction-collision geochemical properties. This suggests their origin at the early collision stages and dates this subterrain formation not older than 45 Ma.

[53] The *Tunaicha Subterrain* has a slab-overthrust and north-dipping nappe-fold structure and is composed of slab and nappe packages, where the Mesozoic oceanic and Late Cretaceous-Paleocene rock complexes are superposed (Figure 5, II–II and III–III). The nappes are separated by the zones of terrigenous and, occasionally, serpentinite melange and were deformed in the southern part of this subterrain by the formation of tectonic windows, where the subarkose turbidites of the Gorbusha rock sequence, or the melanged rocks of the Tonin-Aniva Terrain, are exposed (Figures 4 and 5). The Tunaicha Subterrain includes the Yunona and Kazachka packages of slabs and nappes, which occupy the frontal and rear positions, respectively.

[54] The articulation of the Chaika and Tunaicha subterrains was mapped in the area southeast of the Tunaicha Lake, being hidden, to a great extent, under the Cenozoic sediments. The northern part of the Chaika Subterrain is deformed by a system of sublatitudinal reverse faults and strike-slip thrust faults, which followed the structural style of the Tunaicha Subterrain. The upper, oceanic and island-arc, slabs of the Chaika subterrain are curved to the northeast, whereas the lower Gorbusha slab is turned, together with the Vavai melange, under the Tunaicha Subterrain, residing in its basement. The northwestern flank of the Okhotsk Massif is most highly eroded and is cut off in the north by the NE-striking faults. North of these faults, the rocks of the Tunaicha Subterrain do not show any indications of hornfels development, the granitoid boulders being restricted to the Oligocene-Lower Miocene conglomerates of the neoautochthon, and being absent in its Upper Eocene basal conglomerates. This suggests that the structural connection of these subterrains might have taken place at the Eocene-Oligocene boundary or during the Early Oligocene with the displacements of the granitoids to the surface. This agrees with the absence of the Late Eocene neoautochthon in the junction zone of these terrains, which can be explained by its erosion.

[55] **The Susunai Terrain** is composed of metamorphic and metamorphosed rocks, which differ in their composition, age, structure, and metamorphic grade in the eastern and western parts of the terrain, known as the West Susunai and East Susunai subterrains (Figures 3 and 6). The southern and western spurs of the ridge show outcropping phyllites grading to unmetamorphozed terrigenous rock sequences. They are interpreted as the eastern flank of the West Sakhalin Terrain.

[56] **The West Susunai Subterrain** is the zone of moderate to high-pressure metamorphism, which differs from the adjacent West Sakhalin Terrain and from the East Susunai Subterrain by its higher metamorphism. The metamorphic rocks were classified into the Late Cretaceous Krasnoselskaya and Mesozoic Sokolskaya metamorphic rock series. The rocks of both series experienced retrograde metamorphism in low-pressure greenschist facies.

[57] This subterrain is basically composed of black plicated, as well as green and blue, schists, with the high predominance of the former, all of them being known as the Krasnoselskaya metamorphic rock series. These rocks contain abundant Na-amphibole of the crossite-riebeckite series and some local Na-pyroxene and lawsonite. The blue schists trace the synmetamorphic faults and are accompanied by small serpentinite and talcite protrusions. As follows from their petrogeochemical characteristics, the greenschists developed mainly after the oceanic intraplate subalkaline basalts and, to a lesser extent, after the MOR Tholeiite basalts.

[58] The riebeckite schists showed their ages to be 67, 77, and 85 million years. Proceeding from the Early Cretaceous-Cenomanian age of the initial rocks for the black shale (spores, pollen, and radiolarians) and from the Paleocene-Eocene age of their diaphthoresis (see below), the formation of these metamorphic rocks was dated Late Cretaceous.

[59] The *Sokolskaya metamorphic rock series* is represented by the amphibolite composing slabs in the serpentinite melange and in the areas underlain by the rocks of the Krasnoselskaya metamorphic rock series, and also by the lawsonite-glaucophane schists and eclogite-like metasomatic rocks developed after amphibolite. As follows from the petrogeochemical data available, the protoliths of the amphibolite were the alkaline and subalkaline igneous rocks of interoceanic rises and more rarely MOR basalts.

[60] The glaucophanized amphibolites and eclogite-like rocks were K-Ar dated 133–135 million years, whereas the amphibolite showed an older 206 Ma age [*Egorov,* 1969]. The amphibole-bearing schists were dated 90–92 Ma and seem to have formed after amphibolites were included into the late Cretaceous Subterrain.

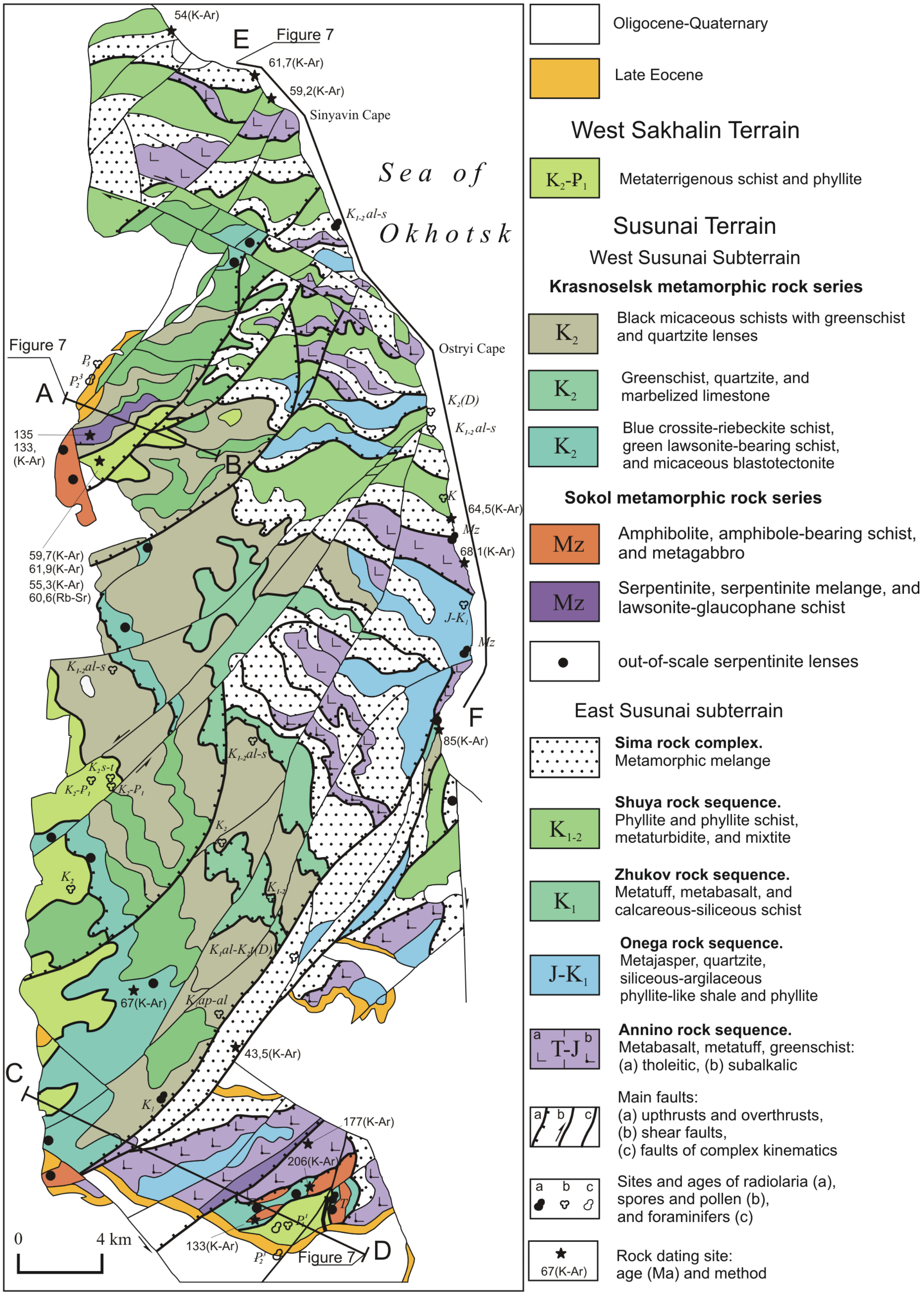

Figure 6. Schematic geological map of the Susunai Range (Susunai Terrain).

[61] **The East Susunai Subterrain** is composed of the irregularly metamorphosed and tectonically layered Annino, Onega, Zhukovka, and Shuya rock sequences. The stratigraphic contacts were found only between the Annino and Onega rock sequences.

[62] The metabasalts of the Annino rock sequence occur as sheets at the bases of all tectonic rock packages and as slabs in the melange zones. These rock sequences were classified into two types. The rocks of one of them are developed in the central part of the subterrain and are represented by the alternation of MOR Tholeiite metabasalts and metahyaloclastites with scarce quartzite lenses (Figure 6). The rocks of the second type are developed in the southern part of this subterrain and are represented by intraplate subalkaline and alkaline intraplate metabasalts, hyaloclastites, and tuff breccias with microquartzite layers and occasional limestone lenses. These rocks are not more than 700 m thick. They were dated Triassic-Jurassic proceeding from the K-Ar dating of the metabasalts (177–179 Ma) and from the age of the overlying Onega rocks.

[63] The *Onega rock sequence* is represented by microquartzite layers and quartzite-like schists, developed after siliceous and siliceous-argillaceous rocks, which are replaced upward by green and black paramorphic schists developed after tuff, tuffaceous siltstone, and silty pelite. This rock sequence is more than 700 m thick. The rocks were dated Jurassic-Lower Cretaceous proceeding from the scarce radiolarians found in the siliceous rocks and from the spore and pollen remains found in the black paraschists.

[64] The *Zhukovka rock sequence* is widespread in the central part of the terrain, along its contact with the West Susunai Subterrain. This rock sequence is represented by the coarse alternation of siliceous shale and tuffaceous silicite, varicolored metahyaloclastite and phyllite developed after the tuffite and tuffaceous siltstone, and slabs of poorly metamorphic subalkaline metabasalts of intraplate origin. The petrogeochemistry of the metabasalts and the total appearance of this rock sequence make it similar to the Lower Cretaceous hard rock sequences of the Tonin-Aniva Terrain. This rock sequence has a maximum thickness of 970 m. The siliceous shale was found to contain Cretaceous spores and pollen. This rock sequence was dated Early Cretaceous.

[65] The *Shuya rock sequence* is composed of alternating phyllites, and phyllitic sandstones with lenses of quartz-like schist, quartzite, green para- and orthoschists, and limestone. Some of these rock sequences show a chaotic structure and seem to be of olistostrome origin. The quartzite-like schists originated after high-quartz sandstones, whereas the metajaspers and cryptogranular quartzite, found in the Annino and Onega rock sequences, are not characteristic here. Found in the schists were Albian-Late Cretaceous palynological complexes, containing Devonian relict spores.

[66] The structure of the Susunai Terrain varies regularly from the older metamorphic rocks to the young ones. The early metamorphic structure of the West Susunai Subterrain has a submeridional strike, dips to the west, and is deformed to large flexure-type folds, the antiforms of which show that the serpentinite melange with amphibolite slabs had been squeezed out from them, whereas the synforms are complicated by the metamorphozed turbidites of the West Sakhalin Terrain (Figure 7). The time of these deformations is limited by the Early Paleocene, as proved by the dating of the West Sakhalin turbidite metamorphism (61.9–59.7 million years, using K-Ar and Rb-Sr dating) and by the age of the sediments overlying these turbidites. The subsequent deformations were associated with the structural formation of the East Susunai Terrain and with the total exhumation of this metamorphic terrain.

[67] The synmetamorphic structure of the East Susunai Subterrain varies in the northern direction from the northeastern one, formed by the tectonic slabs of different dips, to the sublatitudinal one, controlled by the overthrust sheets of southern vergence, and emphasized by the Sima metamorphozed terrigenous melange. The age of this melange becomes younger northward from 64.5 to 54 million years. This subterrain shows at least four structural styles of different ages estimated using the age of its micas. The early structural elements of the NNE strike had originated in the course of the subduction of the volcanic siliceous rocks under the West Susunai Subterrain at the end of the Cretaceous (68 Ma) and of their swell-like folding during the deformation of the Cretaceous subduction zone at the beginning of the Paleocene (64 Ma). The main synmetamorphic sublatitudinal structural style reflects the tectonic stratification of these structural features and their subduction in the northern direction in the Paleocene-Early Eocene (61–54 million years). The late metamorphic structural style was associated with the upthrusting of the whole Susunai Terrain in the SE direction during the Middle-Late Eocene (43.5 Ma).

[68] **The collision-type structural** features originated along the contacts of the terrains discussed above. The typical collision-type structural feature is the articulation zone of the Tonin-Aniva and Ozerskii accretion terrains, represented by the Vavai melange and marked by the Middle Eocene-Early Oligocene syncollision granites (Figure 4).

[69] The *Vavai melange zone* is traceable over a distance of about 70 km, with its width ranging from 5 km to 8 km, from the Sea of Okhotsk coast to the western flank of the Tonin-Aniva Peninsula, where, together with the structural features of the Tonin-Aniva and Ozerskii terrains, it is cut off by the shear faults of the Merei suture zone (Figure 4). Over its total length, the Vavai melange includes tectonic lenses of serpentinite, ophiolite gabbroids, and zones of serpentinite melange, which allow one to interpret it as a suture structural feature. Its southeastern extension is marked by the outcrops of melange rocks with small serpentinite protrusions and gabbroid lenses at the protruding capes of the Sea of Okhotsk coast. The melange includes the slabs of the Middle Cretaceous accretion rocks, the rocks of the Meso-Pacific crust, and the Campanian-Paleocene island-arc rocks.

[70] The structure and structural styles of the Vavai melange in the eastern and western parts of the peninsula differ markedly and generally repeat the structural specifics of the subterrains of the Ozerskii Terrain. The imbricate structure of the Vavai suture zone is overlain, in the northern part of the peninsula, by a Late Eocene-Oligocene neoautochthon. In the south of the peninsula the melange rocks are metamor-

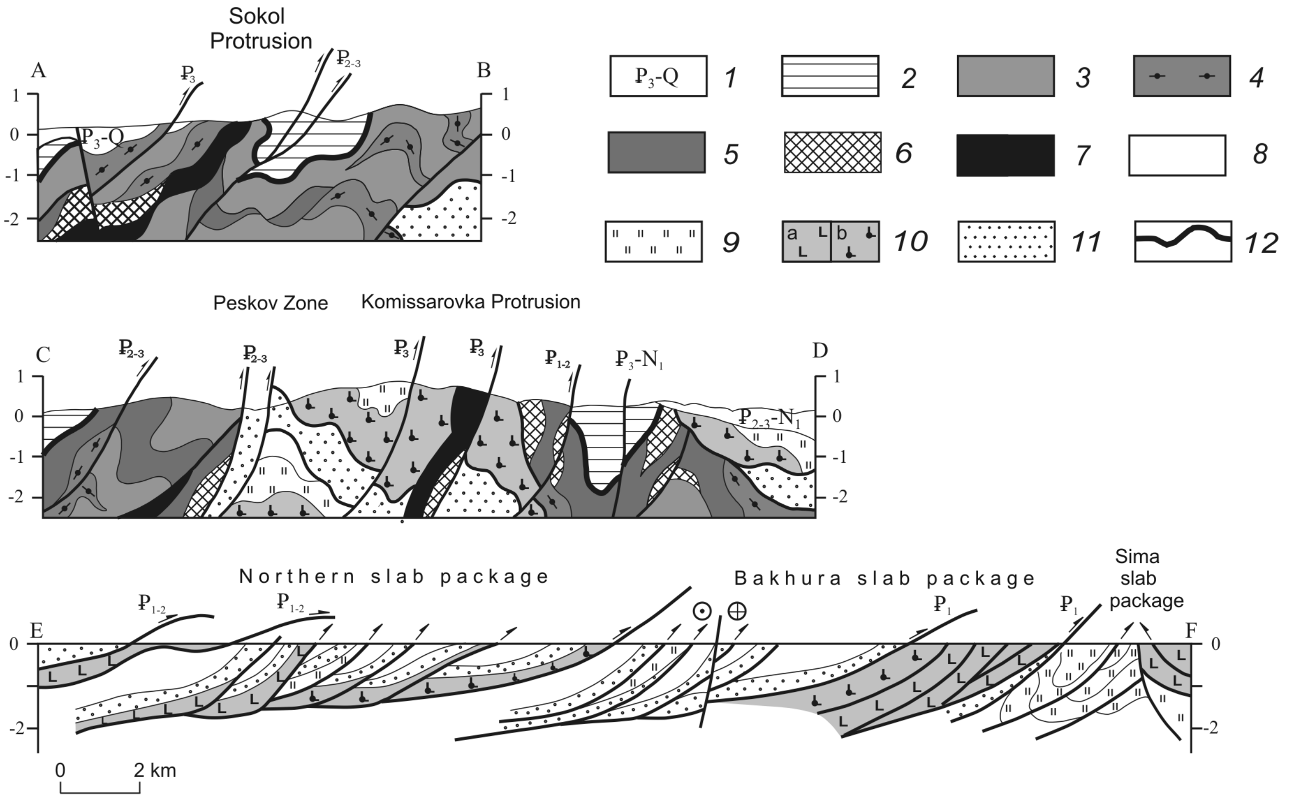

Figure 7. Geological profiles across the Susunai Range. (1) postmetamorphic rocks and their ages; (2) phyllite and phyllite-like schists of the West Sakhalin Terrain (K_{1-2}-P_1^1); (3–12) Susunai Terrain: (3–7) West Susunai subterrain: (3–5) Krasnoselskaya metamorphic rock complexes (K_2): (3) black schist, (4) green schist, (5) blue schist; (6, 7) Sokol metamorphic rock series (J-K_1): (6) amphibolite, (7) serpentinite; (8–12) East Susunai subterrain: (8–11) metamorphic rock sequences: (8) Shuya (K_{1-2}), (9) Onega (J-K_1), and (10) Annino (T-J), metabasalts: (a) tholeitic, (b) subalkaline and alkaline; (11) metamorphic melange; (12) terrain boundary.

phosed to hornfels at the contact with the Middle Eocene-Early Oligocene syncollision granites.

[71] The connection zone of the Susunai and West Sakhalin turbidite terrains can be ranked as the structural features discussed above. The synmetamorphic structure of the West Susunai subterrain was deformed along this zone during the Paleocene with the exhumation of the Early Cretaceous high-pressure schists and the low-pressure greenschist metamorphism of the West Sakhalin turbidites. West of this connection zone, the turbidites are cut by the Early Paleocene rhyolite of subduction-collision origin (Figure 3). These processes were synchronous with the closure of the West Sakhalin trough and operated in the tension field of the left-lateral strike-slip fault, the eastern flank of which was the Merei suture zone.

[72] The *Merei Suture Zone* is one of the key structures of the Southeast Sakhalin, which extends as a band of the highly deformed Cretaceous deposits and melanges, 4–6 km wide, north of the Aniva Gulf coast (Figures 3 and 8). This zone separates the accretion terrains from the continental margin and is interpreted as a shear zone.

[73] The Merei Zone was found to include three lithostratigraphic units, similar to the rocks of the same age in the conjugated terrains, and the zones of shale melange with various tectonic inclusions (Figure 8). The most widespread rock is the Late Cretaceous-Paleocene silty pelite similar to that of the upper rock unit of the Tunaicha Subterrain and to the rocks of the same age in the southern surroundings of the Susunai Ridge. The pre-Turonian rocks are separated into two tuffaceous terrigenous and siliceous-argillaceous rock sequences, developed in the southern and northern zones, respectively, and separated by terrigenous melange zones.

[74] The *Lower Cretaceous tuffaceous-terrigenous rock sequence* is represented by silty pelite with interlayers of highly quartzitic sandstone and silicic argillite and members of volcaniclastic and pyroclastic rocks. This rock sequence is intricately deformed marking a few structural styles, the main of which is represented by the combination of isoclinal and SE-overturned folds with their hinges dipping steeply (20–60°) to the southwest, by ubiquitous cleavage, and by slab-type faults of the NE strike, showing distinct indications of a left-lateral strike-slip fault.

[75] The *Albian-Cenomanian siliceous-argillaceous rock sequence* is composed of siliceous siltstone and argillite with interbeds of siliceous sandstone. These rocks are substituted downward by tuffaceous silicite, their upper parts containing

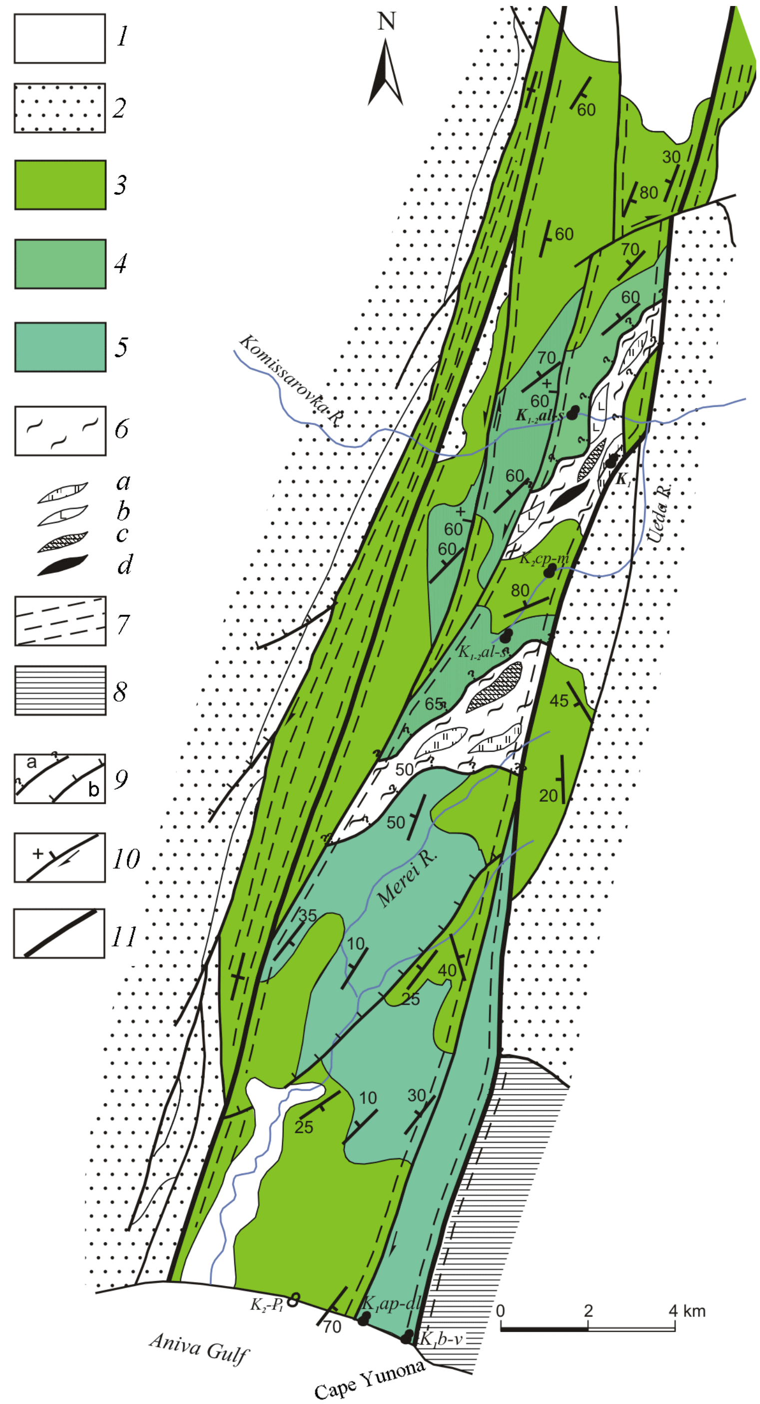

Figure 8. Schematic geological map of the Merei R. and Komissarovka R. basins (Merei suture zone). See Figure 4 for the location of this area. (1) Late Miocene-Quaternary; (2) Late Eocene–Early Miocene; (3) Late Cretaceous-Paleocene siltstone, argillite, siltstone interbedded by sandstones and marl concretions; flyschoid interbedding of pelite, siltstone, and sandstone; (4) Albian-Cenomanian siliceous argilaceous rock sequence of tuffaceous siltstones and sandstones, tuffite, tuff, siliceous siltstone, tuffaceous siltstone members of gradation structure; (5) Lower Cretaceous tuffaceous-terrigenous rocks; (6) shale melange (P_{1-2}) with blocks and slabs of (a) jasper, (b) basalt, (c) gabbro, and (d) serpentinite; (7) mylonite; (8) Ozerskii Terrain; (9) overthrust and reversed faults: (a) pre-Late Eocene, (b) Oligocene post-accretion; (10) shear and upthrust shear faults; (11) faults of complex kinematics.

aleuropelite members with marly sandstone interbeds and marl concretions. In the upper part of this rock sequence, Albian-Cenomanian radiolarians were found in the siliceous sandstones.

[76] The *Late Cretaceous-Paleocene aleuropelite sequence* was found to be represented by two types of the rock sequence. In the western part of the Merei Zone and at the Aniva Gulf coast, this rock sequence is composed of homogeneous silty pelite with occasional interlayers of graywacke sandstone beds. These rocks are transformed to dark-gray clay-like mylonite.

[77] In the eastern part of the Merei Zone this rock sequence rests (supposedly conformably) on the sequence of siliceous-argillaceous rocks. It is also composed of aleuropelite with marl concretions, yet includes the notable number of sandstone and tuffaceous sandstone interlayers, members of alternating normal sedimentary and tuffaceous rocks and occasional siliceous siltstone interbeds.

[78] The western and eastern segments of the Merei Zone differ in structure. The western segment is composed of mylonite bands, 1–2 km wide, developed after the Upper Cretaceous-Paleocene aleurolite of the West Sakhalin Terrain. The eastern segment of this zone is composed of steeply standing tectonic slabs of NE strike, composed of Cretaceous rocks of different formations and of melange, resembling that of the Ozerskii Terrain. Compared to the NE orientation of the slabs, the internal structure of this zone is characterized by the development of pre-Eocene left-lateral structural paragenesises, whereas the right-lateral shears are Neogene ones, have submeridional strikes, and inherited the structural heterogeneities inside and at the flanks of the suture zone. The comparison of the Cenozoic rocks, as well as of the structural and geophysical indications on both sides of the Merei Suture, suggested the magnitudes of the right-lateral movements to be 35–40 km.

Tectonic Models for the Terrain Formation

[79] The results of the terrain analysis of the southeast Sakhalin and the data available for the West Sakhalin, Moneron, and Kem terrains [*Golozubov,* 2004; *Governmental Geological Map,* 2001; *Malinovskii et al.,* 2002; *Piskunov and Khvedchuk,* 1976; *Richter,* 1986; *Zyabrev,* 1992], and for the correlative tectonic units of Hokkaido [*Isozaki,* 1996; *Kiminami et al.,* 1992; *Kiyokawa,* 1992; *Niida and Kito,* 1986, to name but a few], suggest that the accretion-type structural features of the Hokkaido-Sakhalin fold system had formed, at least, from the Aptian to the Eocene (Figure 9).

[80] Before describing the tectonic evolution of the region, I will first discuss the models of the South Sakhalin terrain formation and the basic processes and mechanisms of the growth of the Cretaceous-Early Paleogene East Asian margin. I believe it more expedient to begin my review with the Susunai Terrain which occupies the internal position in the assemblage of the South Sakhalin accretion-type structural features and restricts, together with the Kamuikotan Terrain, the propagation of the Cretaceous-Paleogene accretion-type rock complexes from the west.

[81] **A model for the formation of the Susunai terrain.** The models describing the origin of the early metamorphic structure and the formation of early (pre-Late Cretaceous) metamorphic rock complexes were offered by Japanese geologists for the Kamuikotan Terrain [*Sakakibara and Ota,* 1994]. Like in the Susunai Terrain, in the Kamuikotan Terrain the protoliths of the amphibolite were the pre-Late Jurassic ophiolites of the Sorachi Terrain [*Dobretsov et al.,* 1994; *Komatsu et al.,* 1992], which are believed to have been formed in the environment of the large intraoceanic Early Mesozoic Mikabu-Sanbosan-Sorachi Plateau (which is here referred to as the Sorachi Plateau) [*Kimura et al.,* 1994; *Maruyama,* 1997]. The datings of the amphibolites from South and Central Sakhalin and Hokkaido fit in the time interval of 206–145 Ma [*Dobretsov et al.,* 1994; *Egorov,* 1969; *Governmental Geological Map...,* 2001; *Khanchuk et al.,* 1988; *Komatsu et al.,* 1992]. These data suggest that the metamorphic rocks of the epidote-amphibolite facies had accumulated over the larger time interval of the Triassic and Jurassic during the growth of the Sorachi oceanic plateau and the tectonic layering of its basement under the conditions of high-pressure (Figure 10a). The amphibolites experienced Early Cretaceons (135–120 Ma) lawsonite-glaucophane metamorphism wich resulted in the formation of the metamorphic rocks of metasomatic origin, widely varying from blueschists to eclogite like metasomatic rocks. As follows from the isotopic, geochemical, and petrologic studies of the rocks, the metaophiolites had been glaucophanized under the condition of high-pressure dynamic metamorphism at the depths of 25–30 km during the rapid subduction of the oceanic plate (apparently, the Izanagi Plate) with the tectonic layering and underplating of the subducted fragments of the Sorachi Plateau (Figure 10b).

[82] Both in the Susunai and Kamuikotan terrains, the metaophiolites and Early Cretaceous high-pressure rocks were included into the Late Cretaceous metamorphic rocks and serpentinite melange as allochthonous slabs. However, they occur as large slabs in the southern segment of the Kamuikotan terrain, and merely as small tectonic slabs, subordinate to the serpentinite melange and Late Cretaceous metamorphic rocks, in the northern segment of this and also in the Susunai terrain. This suggests that the emplacements of the pre-late Cretaceous metamorphic rocks into the later accretion-subduction structural features was accompanied by their breaking into individual slabs and by their spreading in the northern direction under the conditions of left-lateral oblique subduction or transform fault movements along the continental margin.

[83] The Kamuikotan Terrain is known for its abundant Albian-Early Cenomanian metamorphic rocks (110–95 Ma). The predominant rock is apoterrigenous shale, whereas the green and blue shales, quartzite, and metaophiolite composing a smaller number of the tectonic slabs are included into the black shale matrix [*Ishizuka et al.,* 1983; *Komatsu et al.,* 1992; *Ota et al.,* 1993]. The regular patterns of this kind are usually interpreted as the involvement of the large volumes of turbidites filling the trench and composing the basal parts of the accretionary prisms. The latter include the tectonically layered fragments of the subducted oceanic plate and

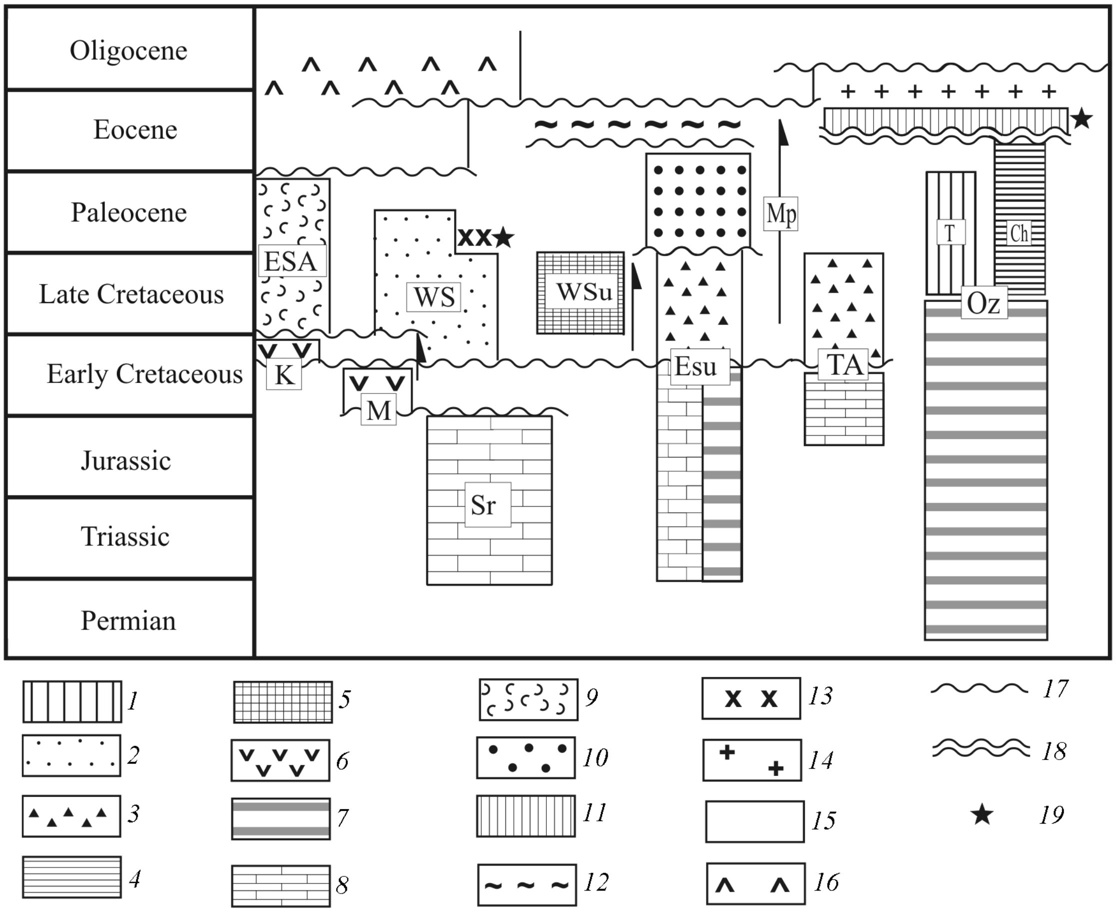

Figure 9. Space-time diagram for the South Sakhalin terrains. Structure-rock complexes: (1) Marginal sea basins; (2) Turbidite basin; (3) Continent-bordering accretion prism; (4) Intraoceanic accretion prism; (5) Subduction zone; (6) Island-arc; (7) Abyssal oceanic plate; (8) Oceanic rise; (9) Continental margin volcanic belt; (10) Sima melange (Ꝑ$_{1-2}$) (11) Vavai melange (Ꝑ$_2$); (12) Peskov Zone melange (Ꝑ$_2^2$). Suturing rocks: (13) rhyolite (Ꝑ$_1$); (14) granite (Ꝑ$_{2-3}$). Overlying rocks: (15) Marine sedimentary and subcontinental rocks (Ꝑ$_{2-3}$-N$_1$); (16) Volcanogenic sediments (Ꝑ$_{2-3}$). (17) Accretion onset; (18) Amalgamation; (19) Collision. Abbreviation: SO – Sorachi Terrain, K – Kem Terrain, WSu and ESu – West- and East Susunai subterrains, Tu and Ch – Tunaicha and Chaika subterrains, ESA – East Sikhote Alin volcanic belt. See Figure 2 for the legend.

of the oceanic crust sediments and had experienced progressive high-pressure metamorphism beginning from the depths of 10–15 km (Figure 10c). Similar rocks seem to have been developed in the Susunai Terrain, as evidenced by singular datings of 90 Ma and 92 Ma [*Egorov,* 1969] obtained for amphibole-mica schists, yet, without associating them with any particular structural features. The latter circumstance does not allow one to identify them as an independent rock complex and to use them in geodynamic analysis.

[84] It can, thus, be concluded that at the end of the Early to the beginning of the Late Cretaceous the oblique and still rapid subduction of the Izanagi Plate was accompanied by the subduction and simultaneous spreading of the fragments of the oceanic plateau along the convergence contact. Because of their sizes and lower density, compared to that of the oceanic lithosphere, the previously subducted plateau blocks might be able to block the further subduction in the southern segment of the subduction zone. In addition, *Ohta* [1996] suggested that the floating up of these blocks might have caused the formation of a mantle wedge and the rising of the old accretion-type structural features of the continental margin (Figure 10c). These processes might have caused the protrusion of the serpentinite melange and metamorphic ophiolites into the accretionary prism and into the basal rocks of the sedimentary rock cover, controlled by the left-lateral displacements in the rear of the accretion prism.

[85] The accretion prism grew in size, both laterally, because of the accumulation of olistostrome rock sequences and the stripping of the hemipelagic sediments from the subduc-

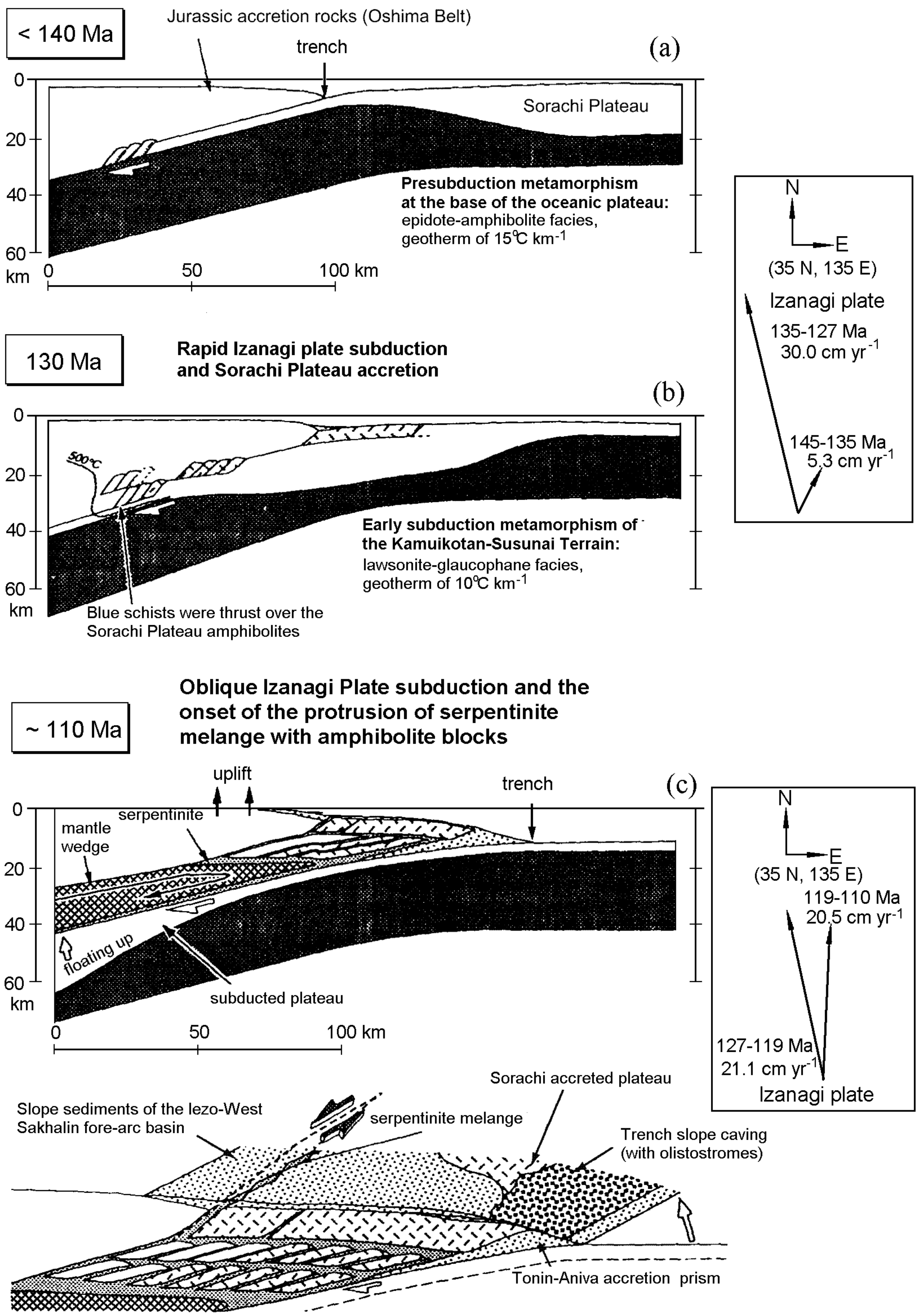

Figure 10. Models for the Early Cretaceous Hokaido-Sakhalin system development demonstrating the origin and evolution of the Kamuikotan and Susunai metamorphic and Tonin-Aniva accretion terrains [after *Ohta,* 1996].

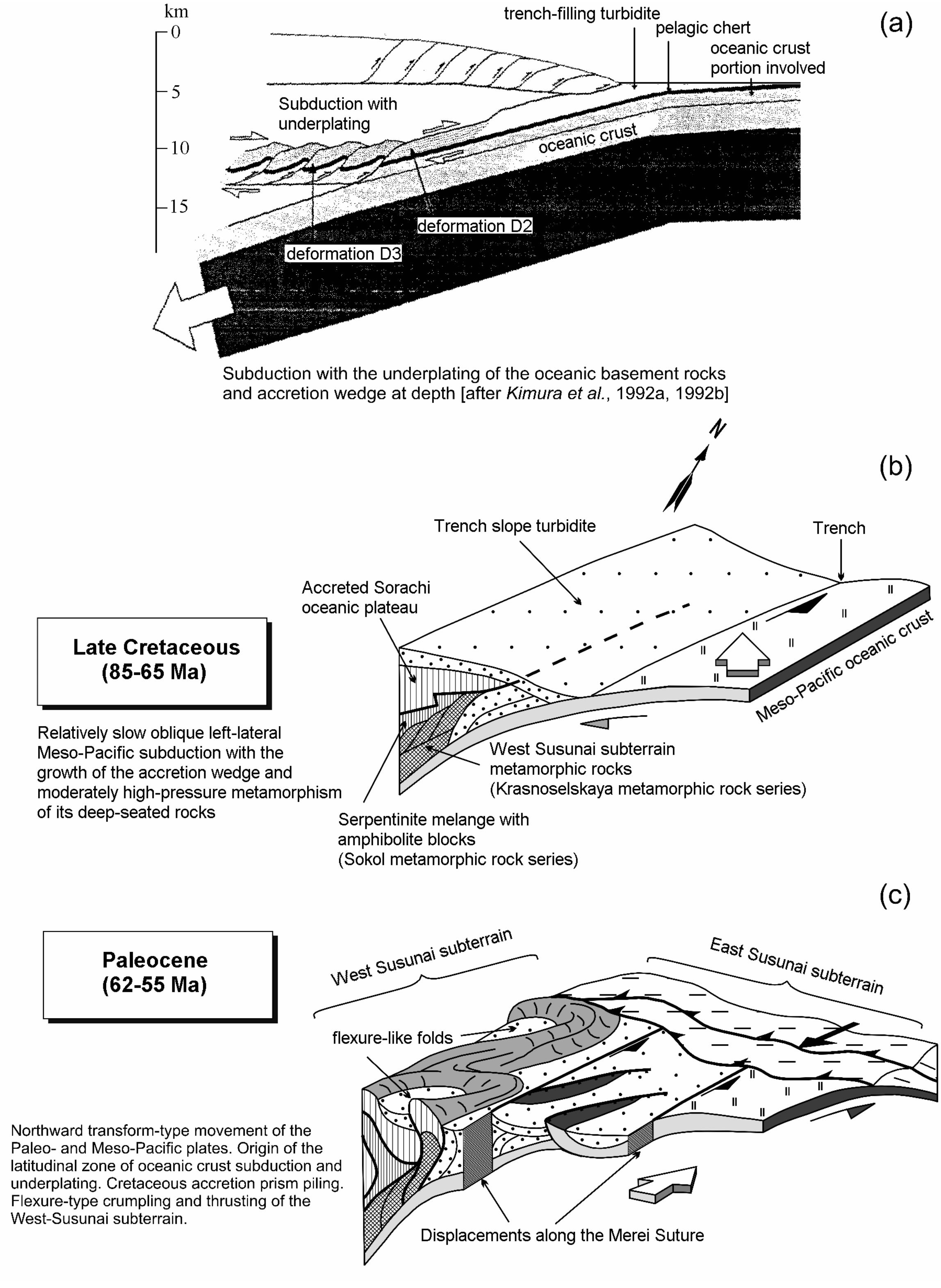

Figure 11. Models for the basal accretion (a) and Late Cretaceous-Paleocene evolution (b, c) of the Susunai Terrain.

tion plate, and vertically, because of the subduction of the accretionary prism, the reworking of the growing volume of the sediments by the basal stripping, the formation of scaly duplexes, and the underplating of the oceanic basement slabs (Figure 11a). Below the depth level of 7–10 km the rocks experienced metamorphic transformations, and the structural pattern of the tectonically layered accretion rock complex acquired the features of a zone of dynamic metamorphic zone. This model was aprobated using the old accretion rock complexes of Japan, Alaska, and Sakhalin [*Kimura et al.,* 1992a, 1992b].

[86] The mechanisms discussed above are typical of the West-Susunai Subterrain, which is interpreted as the deep levels of the Late Cretaceous (85–65 Ma) zone of left-lateral subduction, which developed synchronously with the formation of the West Sakhalin turbidite basin (Figure 11b). The structural pattern of the West Susunai subterrain was shaped as a result of the long-lasting subduction of the Meso-Pacific oceanic rock complexes and of the Middle Cretaceous accretion prism in the northwestern direction. The subducted portions of the accretionary prism experienced dynamic metamorphism of the glaucophane-green schist facies. Its metaophiolites (amphibolite and serpentinite) experienced high-pressure diaphthoresis. The position of the metaophiolites (amphibolite and serpentinite) along the western flank of the subterrain suggests their involvement into the subduction structure during the early phases of its evolution and their subsequent spreading under the conditions of the left lateral subduction. The evidence of the latter is provided by the numerous small serpentinite slabs recorded between the outcrops of the Sokol and Komissarov protrusion zones (Figure 6).

[87] At the Cretaceous-Paleogene boundary the structural features of the Late Cretaceous subduction zone were deformed to the echelon-like system of flexure-type folds with the initial exhumation and the subsequent retrograde metamorphism of the high-pressure rock complexes. These movements were accompanied by the inversion of the eastern segment of the West Sakhalin turbidite basin and by the poor prehnite-pumpelliyte facies metamorphism (K-Ar age of 59.7–61.9 Ma) of its lower parts (Figure 11c). At the western flank of the inversion zone the Campanian-Maestrichtian turbidites were cut by a rhyolite dike belt with the K-Ar age of 68±2 Ma [*Governmental Geological Map,* 2001] of subduction-collision origin (Figures 3 and 9). In the limbs of the flexure-like folds, the high-pressure schists of the West-Susunai Subterrain had been thrust over the poorly metamorphic rocks of the West Sakhalin Terrain. The latter were eroded in the antiform fold hinges, yet, were not eroded in the synforms, being overlain by the Lower Paleocene conglomerates containing West Sakhalin aleuropelite pebbles (Figures 6 and 7).

[88] More eastward, the poorly metamorphozed rocks of the upper structural levels of the Cretaceous subduction zone, classified as those of the East Susunai Subterrain, experienced a swell-like curvature and were collected to NE-trending slabs and folded in the NW vergence with steeply dipping hinges (Komissarov Block and Sima package of slabs). The orientation of the compression structural features and the left-lateral en-echelon type of the flexure-like folds suggest that the rocks were deformed in the tension field of the left-lateral strike-slip fault, the eastern flank of which was the Merei suture. The western flank of the Early Paleogene deformation seems to have inherited the left-lateral strike-slip fault zones along which the metaophiolites and serpentinite melanges were protruded at the boundary between the Early and Late Cretaceous (Figure 10c).

[89] As follows from the structural analysis and isotope dating of the schists (64.5–59.2 million years), the imbricate-overthrust structure of the northern segment of the East Susunai Subterrain was formed during the middle Paleocene, synchronously with the deformation of the Cretaceous subduction zone (Figure 11c). It is interpreted as the north-trending zone of the Paleocene-Early Eocene subduction of the epi-oceanic marginal sea, the mid-Cretaceous accretion prism, and the distal parts of the turbidite basin, and is believed to have been associated with a change in the direction of the paleo-oceanic plate movement from the northwestern (314°) to the NNW one (358–338°). The structural features of the Cretaceous subduction zone and of the metamorphism, involved into the zone of the Late Paleocene-Early Eocene subduction, were tectonically layered by imbricate overthrusts and terrigenous melanges of southern vergence and superposed with the poorly metamorphic Mesozoic rocks of the East-Susunai Subterrain (Zhukovka, Sima, and Bakhura slab packages).

[90] To sum up, the northern part of the Susunai Terrain is distinguished by the superposition of the Late Paleocene-Early Eocene sublatitudinal subduction zone over the Cretaceous submeridional, intricately deformed zone of high-pressure dynamic metamorphism and corresponds to the end-to-end junction of the structural features of the Cretaceous and Early Paleogene subduction zones.

[91] The late metamorphic structure of the Susunai Terrain was formed during the Middle Eocene under the conditions of the single-trend compression from the northwest and reflects the mature exhumation stage of the Susunai metamorphic rocks.

[92] **A Model for the Origin of the Aniva Composite Terrain.** The Aniva composite terrain originated as the structural unit of the Hokkaido-Sakhalin Fold System in the Middle Eocene. The structural features and the great diversity of the tectono-stratigraphic units composing its Tonin-Aniva and Ozerskii terrains suggest their separate formation up to the middle of the Paleogene. The models of the formation of these terrains are presented as paleotectonic profiles in Figure 12.

[93] The Aptian-Cenomanian stage of the Tonin-Aniva Terrain formation agrees with the model of the Middle Cretaceous evolution of the Susunai Terrain (Figure 10c). The accretion structure of this terrain was formed in the course of the left-lateral subduction of some large intraoceanic volcanic rises, accreted to the East-Asia margin during the second half of the Early Cretaceous (Figure 12a).

[94] The rocks of the Aptian-Cenomanian accretion complex (Utesna rock sequence) accumulated during the rapid rising of the continental margin, which was accompanied by the breakdown of the continental slope, avalanche-like sedimentation, and the formation of thick olistostrome and turbidite sequences. The olistostromes were saturated mainly with the clastics of the accreted volcanic plateau and with the products of the synchronous alkali-basaltic volcanism, whereas the turbidite rock sequences with their subarkose clastic material accumulated at the expense of the erosion of the continental land. The fact that the olistostromes con-

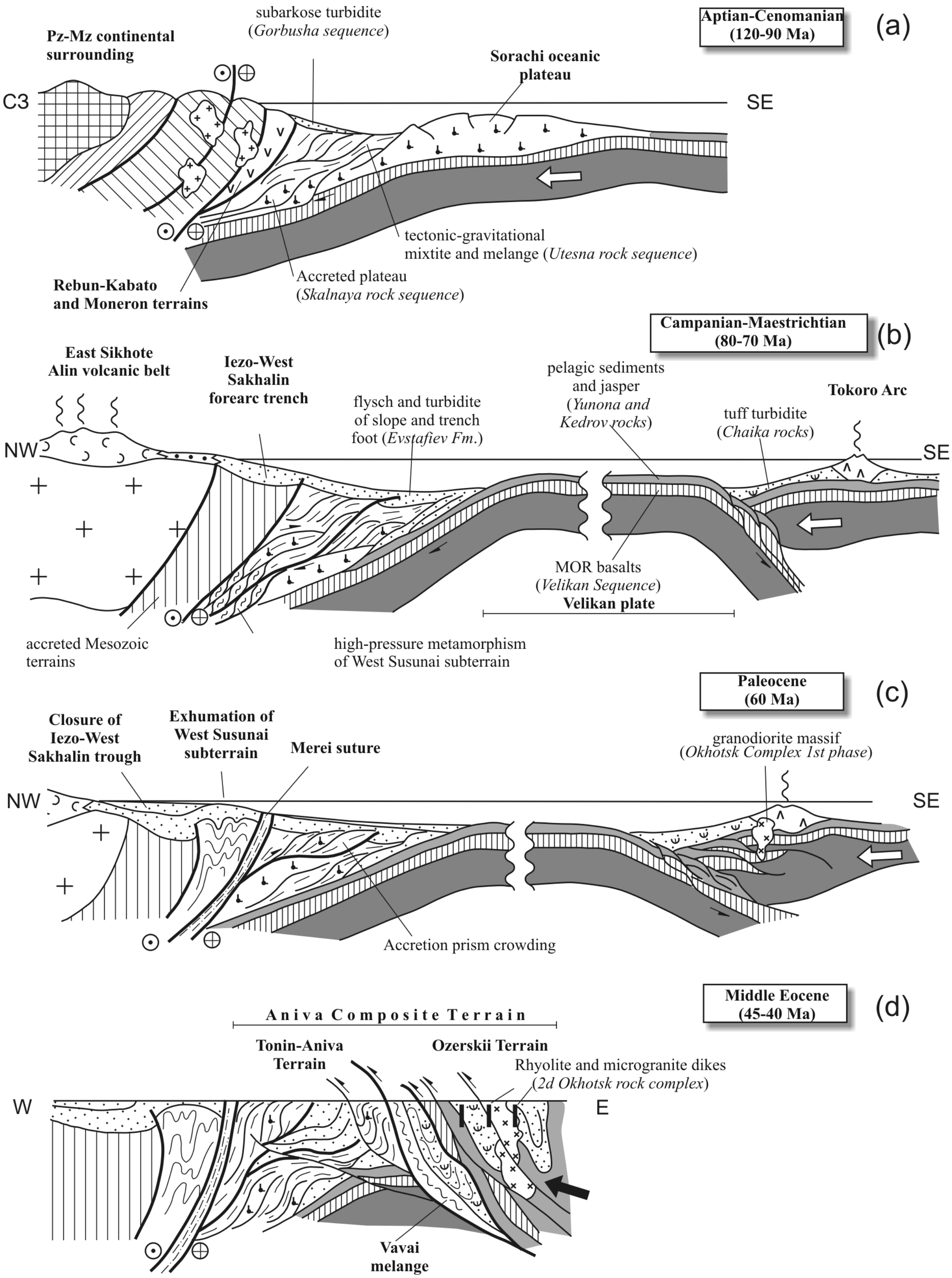

Figure 12. Model for the Aniva composite terrain formation.

tain syngenetic quartz feldspar and high-quartz sandstones, on the one hand, and the Late Permian limestones of rift origin, on the other, suggests that the Utesna rocks had accumulated in the vicinity of the old accretion-type continental margin including continental terrains (Hida, South Kitakami, and others), and the pre-Cretaceous accretion prism terrains with the fragments of Permian intraoceanic plateaus. For instance, the fragments of the Maizuru Plateau are included into the Jurassic accretion-type Honshu and SW Hokkaido terrains, such as, Mino-Tamba, North Kitakami, and Oshima [*Isozaki,* 1996; *Kiminami et al.,* 1992; *Kimura et al.,* 1994; *Maruyama,* 1997; *Maruyama et al.,* 1997]. The potential sources of the Permian continental and exotic clastic material are located in the modern structure of the continental margin at least 600–800 km south of the Tonin-Aniva Terrain position. This suggests that this terrain began to form in lower latitudes.

[95] The subsequent evolution of the accretion prism in the late Cretaceous was synchronous with the accumulation of the Late Cretaceous rocks of the Evstafiev Formation, which covered the Middle Cretaceous accretion wedge and the sediments of the oceanic plate by distal-facies turbidites. The evolution of the Tonin-Aniva Terrain in the conditions of the eastward progradation of the forearc trough of the East Sikhote Alin volcanic belt was accompanied by the displacement of the Middle Cretaceous accretion rock complex to the north along the continental margin, and by its simultaneous crowding and thrusting over the Late Cretaceous turbidite. The movements of this terrain seem to have been controlled by the left-lateral shear zones which inherited the old convergent boundaries and later bounded the Idonnappu and Merei suture-type structural features.

[96] The specific structure of the Ozerskii Terrain explains the accumulation of its rock complexes in the environment of the ensimatic island arc (the Chaika rock sequence and the Okhotsk rock complex) and in that of epicontinental marginal sea (Upper Cretaceous-Lower Paleogene silty pelite). The Gorbusha turbidites seems to have accumulated in the same environment as the Utesna turbidites, yet, in more close association with the continental terrains which supplied quartz-feldspar and granite clastic material (Figure 12a). The presence of the Gorbusha subarkose turbidites in the para-autochthon of the oceanic rock complex can be explained by some specific mechanisms of the tectonic combination of the Ozerskii and Tonin-Aniva terrains.

[97] The Ozerskii Terrain is composed mostly of the Paleo- and Meso-Pacific oceanic rocks which had been emplaced prior to the Campanian in the low-latitude areas, as proved by the finds of Jurassic-Early Cretaceous heat-loving radiolaria, typical of the Tethyan Provinces of the Pacific and of the Alpine Mediterranean belt, by the finds of the Cretaceous Tethyan rudists in the Kedrov rock sequence, and confirmed by paleomagnetic data [*Bazhenov et al.,* 2001, 2002].

[98] The Upper Permian-Middle Triassic metabasalts (Velikan sequence) correspond to the upper part of the second layer of the oceanic crust which had been formed in the environment of a mid-oceanic ridge. The Middle Triassic–Jurassic condensed jaspers of the Yunona rock sequence, overlying the basalts, suggest the environment of a deep-sea abyssal plain. The overlying Upper Tithonian-Lower Cretaceous Kedrov rocks sequence is saturated with pyroclastic material and encloses subalkaline basalt lavas. This suggests the replacement of the deep-sea siliceous sedimentation by the environment of oceanic intraplate volcanism.

[99] The accretion rocks of the Chaika Subterrain accumulated under the conditions of deficient sediments in the forearc basin and in the trench (not more than 1 km thick). This facilitated the deep propagation of the basal breaks into the subducted oceanic crust and the substantial vertical accretion with the underplating of large oceanic crust slabs. The upper structural levels of the accretion complex of the Chaika Subterrain are composed of Campanian-Paleocene tuffaceous turbidite of the Chaika rock sequence and are intruded by Paleocene-Early Eocene diorite-granodiorite bodies of the Okhotsk Complex. The Campanian-Maestrichtian tuffaceous turbidites overlie the Santonian-Early Campanian hemipelagic sediments in the upper part of the oceanic crust, as has been found in the Chaika slab. The specific features of the lithology and structure of these rock complexes suggest that island-arc volcanic activity developed in the Campanian-Maestrichtian time, and diorite-granodiorite intrusions of type I were emplaced in the Paleocene-Early Eocene time (Figure 12c).

[100] As follows from the paleomagnetic data available, the Campanian-Maestrichtian tuffaceous turbidites of the Chaika rock sequence had accumulated at the paleolatitude of 26.6±5.2 N, or in the area about 3000 km south of their modern position [*Bazhenov et al.,* 2001, 2002]. Similar results were obtained for the Tokoro tuffaceous turbidite [*Kanamatsu et al.,* 1992]. This allowed us to classify these rock complexes as the constituents of the same ensimatic arc, referred to as the Tokoro arc [*Bazhenov et al.,* 2001]. As follows from our kinematic analysis, this island arc developed at the leading edge of the Pacific plate and, consequently, the subduction was directed to the ocean. The Tokoro island arc was active up to the Middle Eocene, which is proved by the presence of Early Eocene turbidites in the accretionary rock complex of the Tokoro terrain [*Kiminami et al.,* 1992].

[101] Earlier, *Bazhenov et al.* [2001] reconstructed the system of the Late Cretaceous ensimatic arcs, which separated, during the Campanian, the segment of the Pacific Plate, which later became an epioceanic marginal sea and has been classified as the Velikan Plate in this study (see Figure 12b). The fragments of its oceanic crust and of the overlying Lower Paleogene sedimentary cover are combined tectonically in the Tunaicha Subterrain.

[102] To sum up, the Ozerskii Subterrain was produced by the accretion of the rocks of the oceanic plate which was transformed at the end of the Cretaceous to an epioceanic marginal sea basin, yet, following different mechanisms of its formation in the Chaika and Tunaicha subterrains. The Chaika Subterrain developed as a Campanian-Early Eocene accretion prism of the ensimatic island arc with the subduction of the oceanic plate in the ESE direction, and the Tunaicha Subterrain developed as a fragment of the epioceanic marginal sea crust, accreted to the continental margin in the Paleogene.

[103] The final stage of the formation of the Tonin-Aniva and Ozerskii terrains was associated with their combination

(amalgamation) as a result of the collision of the Tokoro Arc with the continental margin and of the complete absorption of the marginal sea plate. The convergence of the terrains was accompanied by the formation of the Vavai melange zone, which happened to include the slabs of the near-continent accretion prism, oceanic crust, and island-arcs (Figure 12d). The interaction of the terrains followed the mechanism described in the model of arc-continent collision [*Konstantinovskaya,* 1999, 2003]. The structural features of the Tonin-Aniva Terrain were overthrust in the western direction. The eastern flank of the terrain experienced crowding and was transformed to the East Aniva slab-type overthrust zone.

[104] During the early collision stage the accretion rock complex of the Chaika Subterrain experienced compression in the SE-NW direction and crowding along the NW-trending overthrusts. The forearc basin of the island arc was basically demolished, being preserved only as allochthonous slabs in the boundary-type Vavai melange or as the fragments bordering the igneous rock arc. That period of time seems to have been marked by the emplacement of the dikes of the second phase of the Sea of Okhotsk Complex. The late collision stage was marked by sublatitudinal compression, obviously associated with the beginning of the Pacific motion in the WNW direction 43 million years ago [*Engebretson et al.,* 1985]. The structural features of the Chaika Subterrain and Vavai melange acquired the western vergence, and the intrusion of the collision granites of the Aniva Complex terminated the process of the terrain assemblage.

[105] The Aniva composite terrain, which was formed after the terrain amalgamation, moved to the north along the Merei shear zone up to its final accretion to the continental margin at the end of the Eocene. The northern part of this composite terrain was included into the Paleogene subduction zone with the tectonic layering and subduction of the crust of the epioceanic marginal sea (Tunaicha Subterrain) and of the Tonin-Aniva terrain structural features (Figure 5, Profile II–II), as well as with the transverse wedge-shaped curving of the Chaika Subterrain structural features (Figure 4).

[106] The syncontinental structural features of the Hokkaido-Sakhalin fold system form the Moneron and West Sakhalin terrains, the formation models of which are still a matter of discussion.

[107] As follows from the modern views [*Malinovskii et al.,* 2002; *Rozhdestvenskii,* 1993; *Simanenko,* 1986], the volcanic rocks of the Moneron and Rebun-Kobato terrains are interpreted as the rock complexes of the outer ocean-bordering part of the Early Cretaceous Moneron-Samarga island arc. The volcanogenic sediments of the East Sikhote Alin, interpreted as the Kem Terrain [*Golozubov,* 2004; *Khanchuk,* 1993; *Malinovskii et al.,* 2002], were ranked as the rear part of the arc and as the back-arc basin. It is believed that after the accretion of the arc and the closure of the back-up basin at the boundary between the Early and Late Cretaceous, their rock complexes were covered by the rocks of the East Sikhote Alin volcanic belt. The terrigenous rocks of the Iezo and West Sakhalin terrains are treated, in terms of this view, as the rocks of the forearc trough, which had accumulated during the Early Cretaceous in front of the island-arc, and as the rocks that had accumulated during the Late Cretaceous-Paleocene, which grade eastward to the Cretaceous-Paleogene accretion-type rocks of Central Hokkaido and East Sakhalin [*Melankholina,* 1988; *Rozhdestvenskii,* 1993; *Zyabrev,* 1992]. The modern positions of the outer and inner parts of the Moneron-Samarga Arc are explained by the shear and pull-apart tectonic movements associated with the opening of spreading basins in the Sea of Japan and in the southern segments of the Tatar Strait.

[108] As follows from the data published in the literature [*Piskunov and Khvedchuk,* 1976], the volcanic rocks of the Moneron Terrain could be layered tectonically into at least four rock sequences ranging from 250–400 m to 1500–1800 m in thickness and from Cretaceous to Paleocene in age, being irregularly distributed over the rock sequence, yet, grouping into four age intervals: 141, 118, 103–86, and 77–59 Ma. The structural and lithologic transformations of the basalts grow more intensive progressively toward the tectonic contacts of the blocks and generally down the rock sequence. This explains the significant scatter of the ages by their rejuvenation as a result of the post magmatic thermal events which seem to have been associated with the tectonic movements of the island-arc terrain. In this case, proceeding from the Upper Cretaceous age of the overlying deposits, it can be assumed that the oldest basalts (141–118 Ma) date the initial age of the rock complex which correlates well with the Berriasian-Barremian age of the volcanic rocks composing the Rebun-Kobato Terrain.

[109] The correlation of the Moneron and Kem terrains is not that easy in connection with the recently published new data [*Malinovskii et al.,* 2002]. The volcanic rocks of these terrains vary in age (Berriasian-Barremian and Late Aptian to Early Albian), whereas the sedimentation model and the arcose composition of the turbidites, underlying the Kem volcanics, contradict the view that an active island arc had existed at that time east of the Kem Terrain.

[110] The model of the formation of the West Sakhalin Terrain was discussed by S. V. Zyabrev. He proved that its Cretaceous rock sequence is composed of classic turbidites which had accumulated in the transit and discharge area of the fan-valley systems of the continental slope, oriented from the west to the east and southeast.

[111] The turbidites had accumulated in a cyclic manner, which is recorded in the vertical sequence of the distal and proximal facies. In the southern part of the West-Sakhalin Trough, the most significant change of the distal sedimentation environment to the proximal one is recorded at the end of the Albian to the beginning of the Cenomanian and was associated with the rapid protrusion of the fan-valley systems to the east. In the Early Turonian they retreated, and merely distal sedimentation environment existed up to the Campanian, which later was substituted again by the proximal one [*Zyabrev,* 1984]. The Late Senonian progradation of the basin correlates well with the development of the East Sikhote Alin volcanic belt and with the displacement of the tectonic and igneous activity toward the ocean, the development of the proximal facies at the boundary between the Early and Late Cretaceous can be explained by the structural rearrangement of the transition zone.

[112] The Aptian-Albian turbidites of the Iezo Terrain and of the southern part of the West Sakhalin Terrain, compared to its northern part, are notably poor in the synchronous pyroclastics, include single thin acid-intermediate tuff layers, yet are enriched in quartz-feldspar grains up to the formation of oligomictic quartz-feldspar sandstone beds. This suggests that during the later half of the Early Cretaceous the southern segment of the Iezo-West Sakhalin basin developed at a significant distance from the region of island-arc volcanic activity, residing, on the contrary, in the region of continental margin erosion. Only the end of the Albian was marked by the significant effects of the rises composed of volcanic rocks (Moneron and other rises), which controlled the development of proximal turbidite facies. However, they supplied basaltic andesite epiclastic material, which proves them to be structural rather than volcanic rises. This agreed with the timing of the tectonic movements of the Rebun-Kobato and Moneron terrains.

[113] As follows from the biogeographic, geochronological, and structural data, the Aptian-Albian period was the time of the large left-lateral movements (up to 1500–2000 km) of the Central Japan ocean-bordering terrains along the Tanakura, Kurosegawa, Ivaizumi, and other transregional shear zones [*Golozubov,* 2004; *Isozaki,* 1996; *Maruyama et al.,* 1997; *Tazawa,* 1993, to name but a few]. It appears that the formation of the Kem Terrain and of the Aptian-Early Cenomanian part of the West Sakhalin Terrain might have been associated with these tectonic movements. One of the potential versions of the formation of the Aptian-Albian alkaline volcanics of the Kem Terrain is their development in the environment of the left-lateral movements of the internal areas of the transition zone, which were lower than those in the perioceanic region. Following this view, the Iezo and West Sakhalin terrains were formed in the Early Cretaceous as a turbidite trough in the Asian margin, similar to the troughs of the Sikhote Alin fold zone, e.g., the Zhuravlev Terrain [*Golozubov and Khanchuk,* 1995]. During the Late Cretaceous, after the accretion of the oceanic (Sorachi) and island-arc (Moneron and Rebun-Kobato) terrains or after their movements along the margin of the large oceanic ridge and after the rearrangement of the transition zone, these terrains began to develop synchronously with the evolution of the East Sikhote Alin volcanic belt as a fore-arc continental margin trough.

[114] Using the models proposed in this study for the formation of the South Sakhalin terrains, we carried out the tectonic reconstruction of the region. Discussed below are the main periods of its geodynamic evolution.

The Cretaceous-Paleogene Geodynamics of South Sakhalin

[115] As follows from our analysis of the South Sakhalin terrains, the rocks composing them had accumulated from the end of the Paleozoic to the middle of the Paleogene, yet the structural shaping of the Hokkaido-Sakhalin fold system began only during the second half of the Early Cretaceous, from the time when the structural features of the continent-ocean transition zone began to form. The history of the tectonic evolution of the region included five structural rearrangements associated with changes in the kinematics of the oceanic plates, with the rearrangement of the continent-ocean boundary, and with the accretion-collision processes that operated in the Asia-Pacific transition zone.

[116] **Triassic-Jurassic.** As follows from most of the paleotectonic reconstructions available, an open oceanic basin had existed during the pre-Cretaceous segment of the geologic record of the East Asia margin, east of the modern outline of the continent [*Khanchuk et al.,* 1988; *Maruyama et al.,* 1997; *Mazarovich and Richter,* 1985; *Melankholina,* 1988; *Parfenov et al.,* 1999; *Rozhdestvenskii,* 1993; *Zonenshain et al.,* 1990, to name but a few]. The limited paleomagnetic data available for the Mesozoic oceanic rocks of Japan and NE Russia [*Sokolov et al.,* 1997; *Zonenshain et al.,* 1990] suggest their accumulation at low equatorial latitudes with their subsequent displacement in the northern direction. This is also proved by the Tethyan complexes of the Triassic, Jurassic, and Early Cretaceous radiolarians found in the siliceous rocks of the Tonin-Aniva and Ozerskii terrains.

[117] As follows from the reconstructions of the oceanic plates [*Engebretson et al.,* 1985; *Maruyama et al.,* 1997], the Farallon-Izanagi spreading range moved along the East Asian margin during the time of 220–180 Ma. Since that time to almost the middle of the Late Cretaceous (85 Ma), the Izanagi plate interacted with the continental margin. During the Triassic and Jurassic a large volcanic oceanic plateau was formed on the Izanagi plate in association with the Central-Pacific plume activity, known as the Sorachi Plateau [*Kimura et al.,* 1994; *Maruyama et al.,* 1997] and as the Okhotsk Plateau [*Bogdanov and Dobretsov,* 2002], and believed to have been associated with the central Pacific plume activity.

[118] **Early Cretaceous.** At the beginning of the Early Cretaceous the Rebun-Moneron island arc was formed, possibly, because of a change in the vector and velocity of the Izanagi plate movement. The position of this island arc is still not clear. Proceeding from the changes in the heavy mineral associations of the Early Cretaceous sedimentary rocks of Sikhote Alin, South Korea, and Southwest Japan, *Nechaev et al.* [1997] assumed that the island arc of that time had formed at the latitude of South Japan and might have extended in the northeastern direction away from the continental margin with the geodynamic environment changing from that of the continental margin to intraoceanic.

[119] The Izanagi plate moved at that time at the highest velocity (30.0 cm year^{-1}) in the northwestern direction (see Figure 10), which caused its rapid orthogonal subduction under the island arc. Its subduction was stopped by the underplating of the large blocks of the above-mentioned oceanic plateau at the end of the Neocomian (possibly during the Hauterivian-Barremian). The high velocity of the Izanagi plate movement (20.7 cm year^{-1}) caused the detachment of the arc rocks from their basement, as well as of the

accretion prism with the fragments of the accreted plateau. All of these rock components continued their passive movement along the continental margin as the constituents of the oceanic plate. This was facilitated by a change in the direction of the Izanagi plate motion from the northwestern to the northern at the beginning of the Aptian without any changes in the velocity of their movement.

[120] As mentioned above, the Aptian-Albian time witnessed the large-scale left-lateral movements of the Meso-Paleozoic terrains of different types along the Asian margin with the formation of a huge collage and the combination of the Tethyan and transitional biogeographical provinces [*Golozubov,* 2004; *Maruyama et al.,* 1997; *Tazava,* 1993]. The shear-type deformations and huge displacements were typical not only of the perioceanic structural features, but also of the Jurassic accretion-type rock complexes, such as the Mino-Tamba, Northern Kitakami, and Oshima ones. The fragments of the Early Cretaceous island-arc system and of the oceanic plateau were spread along the Asian margin over hundreds to a few thousands of kilometers and produced the Rebun-Kobato, Moneron, and Sorachi terrains. The Izanagi Plate seems to have experienced oblique subduction under the continental margin in different segments of its boundary with the latter, or its transform-type movements. Accretion-type rock complexes, similar to those of the Tonin-Aniva Terrain, accumulated in areas of the subduction environment. Turbidites of subarkose composition, might have accumulated in the areas of the transform-type boundary at the expense of the erosion of the Jurassic accretion-type rocks, as well as of the continental terrain deposits. Active shear-type movements continued during the Cenomanian, as the Izanagi Plate moved along the East Asian continental margin. Moving along with it was the oceanic ridge which separated the Izanagi and Pacific oceanic plates (see Figure 13a).

[121] The global left-lateral movements resulted in the formation of an extensive suture which had inherited the Early Cretaceous convergent boundary and could be traced by the fragments of the accretion-type Northern and Southern Chichibu and Kurosegawa terrains and of the subduction-type Mikabu and Sambagawa terrains of Southwest Japan. The granitoid magmatism of the Rioke Belt developed with some delay in the rear of this structural feature [*Isozaki,* 1996; *Miyashiro,* 1975]. Earlier, it was interpreted as a high-temperature and low-pressure metamorphic belt, forming a pair with the Sambagawa high-pressure metamorphic belt. According to this view, the Sambagawa Terrain and the metamorphic accretion-type terrains, conjugated with it, were "pressed" in the zone of transregional strike-slip faults, whereas the Rioke granite belt was formed as a syncollision tectono-magmatic structural feature. This is proved by the intensive transformation of the granitoids to gneiss and by their tectonic position. As follows from [*Maeda and Kagami,* 1996; *Maruyana et al.,* 1997], the formation of the syntectonic granitoids and the low-pressure metamorphism might have been associated also with the development of the Izanagi-Pacific oceanic ridge along the continental margin. As follows from the age of the igneous rocks (100$\pm$5 million years), the peak of this tectonic activity can be placed at the end of the Early Cretaceous.

[122] The left-lateral strike-slip movements involved the whole of the continental margin. Its Jurassic-Early Cretaceous accretion structure experienced repeated lateral doubling with the formation of large saw-shaped protrusions in the NE direction [*Golozubov,* 2004]. One of these protrusions was formed by the Abakuma, North and South Kitakami, and Oshima terrains of Southeast Japan. It can be supposed that the formation of this doubling zone might have been accompanied by the Aptian-Albian alkaline volcanic activity of the Kem Terrain along its continental side. At the same time the Rebun-Kobato, Moneron, and Sorachi terrains were displaced along the oceanic side of the protrusion and accreted at the end of the Early to the beginning of the Late Cretaceous, having occupied the position close to the modern one, adding the abundant volcaniclastic material to the synchronous turbidite sediments of the southern segment of the Iezo-West Sakhalin Trough. The latter originated as a circum-shear turbidite depression which grew larger in the northern direction and overlapped the Sorachi Plateau accreted segments. The eastern segments of this plateau continued to move to the north and were subducted in the oblique subduction zone (Figure 13b).

[123] **Late Cretaceous.** As the Izanagi-Pacific Ridge had moved along the East Asia margin, the Pacific oceanic plate had been interacting with the latter. During the Late Cretaceous the geodynamic environment in the transition zone changed notably, although was still controlled by the oblique subduction of the oceanic plate. The accreted pre-Late Cretaceous terrains were overlain by the East Sikhote Alin continental margin volcanic belt and by the Iezo-West Sakhalin forearc basin grading eastward to a deep-sea trench.

[124] The process of the oblique subduction was accompanied by the left-lateral movements of the Early Cretaceous prism with the subduction and high-pressure metamorphism of the lower structural levels (West Susunai Subterrain). The upper structural levels experienced crowding and were thrust over the fore-arc turbidite and trench sediments. To sum up, the active continental margin of the Andean type existed in the east of Asia during the early half of the Late Cretaceous. Its accretion-type structural features were formed in the environment of the oblique subduction of the Pacific Plate.

[125] The Early Campanian time witnessed the reorganization of the Asia-Pacific ocean convergent boundary. A system of intraoceanic island arcs was formed in the northwestern part of the Meso-Ppacific Ocean at a latitudinal distance of up to 2000–3000 km from the continental margin [*Bazhenov et al.,* 2001, 2002; *Kovalenko and Chernov,* 2003]. The compilation of the paleomagnetic data available for the Campanian-Early Paleogene island-arc rocks of Hokkaido, South Sakhalin, Kamchatka, and Koryakia shows that the island-arc systems had a NE strike, generally orthogonal to the movement of the Pacific Plate. The southwestern flank of the island arc garland, closest to the East Asian continental margin, was occupied by the Tokoro Arc. This arc developed at the leading edge of the Pacific plate which moved, at the end of the Cretaceous, in the NW direction at a rate of 13.1 cm year^{-1} [*Engebretson et al.,* 1985]. The subduction zone was directed to the ocean, and the arc moved closer to the continental margin, together with the

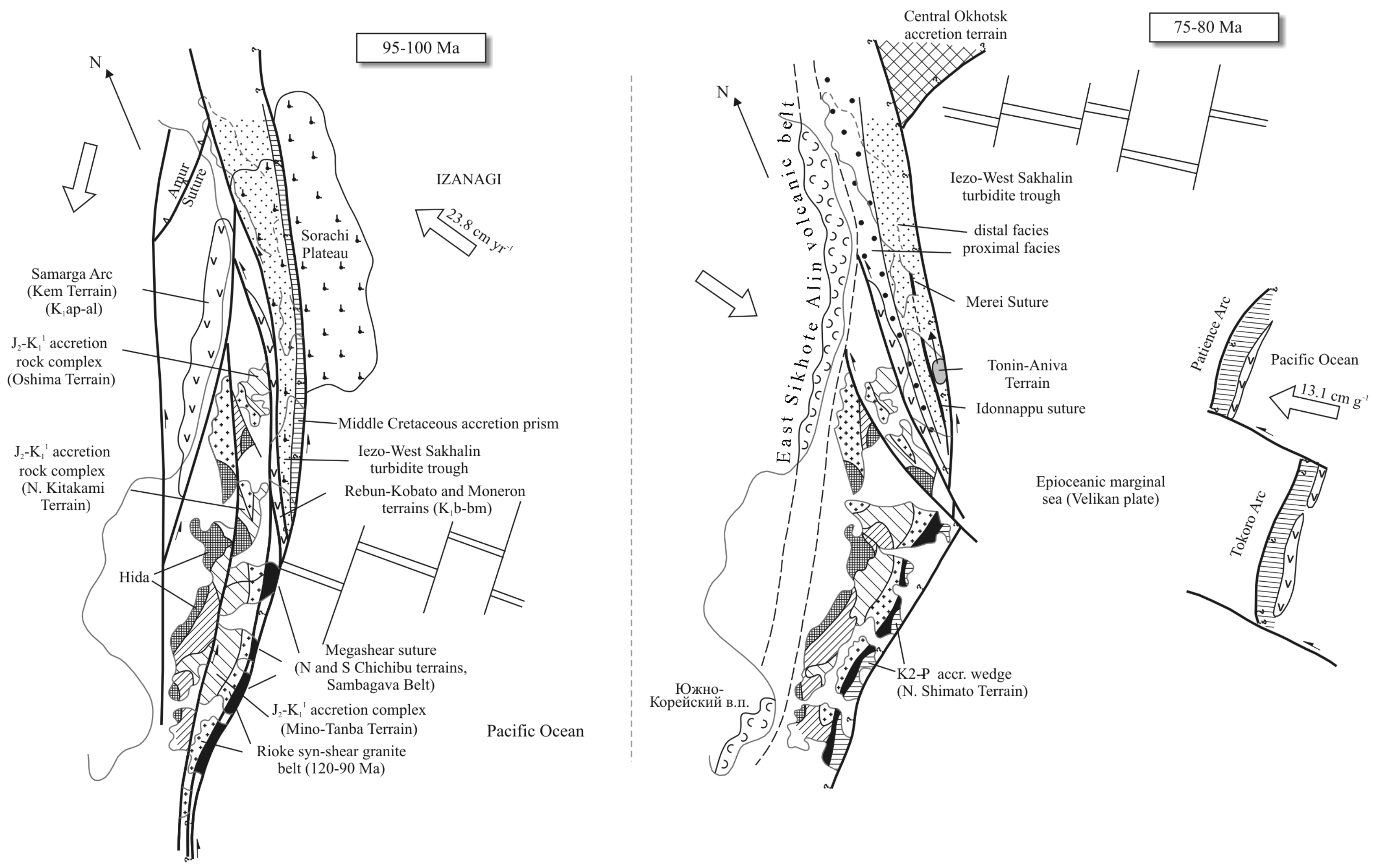

Figure 13. Schematic tectonic reconstruction maps of the East Asia continental margin for the Early-Late Cretaceous. See Figure 1 for the legend.

overhanging oceanic plate. Northeast of the Tokoro Arc, supposedly following the same kinematic model, the East Sakhalin island arc [after *Zonenshain et al.,* 1990], or the Patience island arc [after *Khanchuk,* 1993; *Parfenov et al.,* 1999], was formed and included in the early Paleogene into the accretion-collision structure of the East Sakhalin composite terrain. As to the further reconstruction and substantiation of the early collision of the Patience Arc with the continental margin, we have to admit that this margin had been advanced northwest relative to the Tokoro Arc and joined it along a transform fault (Figure 13b).

[126] The space between the ensimatic arcs and the continental margin was occupied by the microplates of the marginal seas, representing large Paleo-Meso-Pacific segments. The Velikan plate identified in this region (Figure 12b) experienced subduction in the Campanian-Maestrichtian both under the continental margin and under the Tokoro Arc, which contributed to the rapid convergence of these structural features.

[127] To sum up, at the end of the Late Cretaceous the environment of the active Andean-type continental margin coexisted with the environment where the continental-margin and intraoceanic subduction zones existed together with epioceanic marginal seas being located between them. The latter situation has no analogs, yet can be compared conventionally with the modern Polynesian or Philippine Sea situation [*Bazhenov et al.,* 2001, 2002; *Kovalenko and Chernov,* 2003; *Melankholina,* 2000].

[128] **Paleocene.** During the Early Paleogene, the collision of the Patience Arc with the continental margin resulted in the formation of the East Sakhalin composite accretion-collision terrain, and the convergent boundary bordering the continent was reconstructed again. The changes of its configuration were associated with the impossibility of the further oblique subduction in the Central Sakhalin area and with the still acting compression in the northern direction, caused by the movement of the Velikan Plate. These movements resulted in the Early Paleocene transformation of the Late Cretaceous oblique subduction zone to a left-lateral shear zone with the folding of the West Susunai Subterrain and the inversion of the West Sakhalin fore-arc basin (Figure 14a). A sublatitudinal underthrust zone was formed during the Paleocene at the southern flank of the east Sakhalin composite terrain. Subducted along this zone were the marginal marine rocks of the Velikan Plate. The western flank of the subduction zone (East Susunai subterrain) was superposed over the Cretaceous subduction zone producing a trench-trench connection.

[129] It should be noted that almost simultaneously, namely, from the end of the Paleocene to the beginning of the Eocene, the collision of the Achaivayam-Valaga ensimatic island arc with the continental margin rearranged the convergent margin with the formation of a new subduction zone dipping under the continent and extending in the direction of South Sakhalin [*Konstantinovskaya,* 1999, 2004].

[130] To sum up, the beginning of the Paleogene witnessed the extinction of the Cretaceous subduction zone, the migration of the convergence boundary to the south, and the formation of a new Paleogene continental margin, known as the Okhotsk one, at the latitude of the South Sakhalin plate.

[131] The subduction of the Velikan Plate under the Tokoro Arc continued up to the Early Eocene, as indicated by the magmatic activity of the suprasubduction structural features and by the synchronous emplacement of fore-arc tuffaceous turbidite in the Tokoro Terrain [*Kiminami et al.,* 1992].

[132] **Eocene-Oligocene.** The Middle Eocene time witnessed the largest structural rearrangement of the region as a result of the collision of the Tokoro Arc with the continental margin and the closure of the continental-margin epioceanic basin (Figure 14b). The arc-continental margin collision produced different effects and was responsible for the evolution along the collision zone as a result of the oblique convergence of the structural features and of the complex configuration of the continental margin.

[133] In South Sakhalin the collision occurred from the end of the Middle Eocene to the Early Oligocene. Southward, in Hokkaido, the collision of the Tokoro and Hidaka terrains terminated only in the Oligocene, as follows from the ages of the syncollision intrusions [*Maeda et al.,* 1990]. In the southern segments of the collision zone, the accretion-type rocks of the Tokoro and Hidaka terrains experienced progressive metamorphism with the formation of the Hidaka collision-type metamorphic terrain. The oldest synkinematic anatectic granites have been dated Early Eocene (56 Ma, Ar-Ar), [*Komatsu et al.,* 1992; *Osanai et al.,* 1994]. This suggests that the collision might have developed from the south to the north, concordantly with the convergence of the island-arc and continental margin structural features, yet, terminated more rapidly in the north.

[134] The Aniva composite terrain, produced by the amalgamation of the accretion structural features, moved in the northern direction along the Idonnappu and Merei suture zones, and was finally included into the continental margin at the Eocene-Oligocene boundary, or at the beginning of the Oligocene, being located in the junction zone between the Okhotsk and proto-Japanese continental margins. Later the accretion-type structural features of Southeast Sakhalin and East Hokkaido were overlapped by the Late Eocene-Early Miocene neoautochthon.

[135] The proto-Japan continental margin began to form synchronously with or slightly earlier than the Tokoro Arc and continental margin collision, that is, at the end of the Early-Middle Eocene. It was recorded in the slow inherited subsidence along the inverted Cretaceous basin and in the formation of the Middle-Late Eocene zone of calc-alkaline volcanism (proto-Japan volcanic arc), displaced eastward relative to the Cretaceous-Paleocene East Sikhote Alin Belt (Figure 14b). The specific geochemistry of the volcanic rocks suggests their suprasubduction origin and their association with some previously subducted Paleo-Mezopacific slab [*Okamura et al.,* 1998; *Yasnygina,* 2004]. At the back of the volcanic zone, extension began in the later half of the Eocene or at the beginning of the Oligocene, and the proto-Tatar and Central Sea of Japan rift basins began to form, with plateau basalts residing at their continental shoulders. On the contrary, in front of the proto-Japan Arc, en-echelon upthrust-overthrust zones were formed, which marked the axes of the synsedimentation structural highs. The compar-

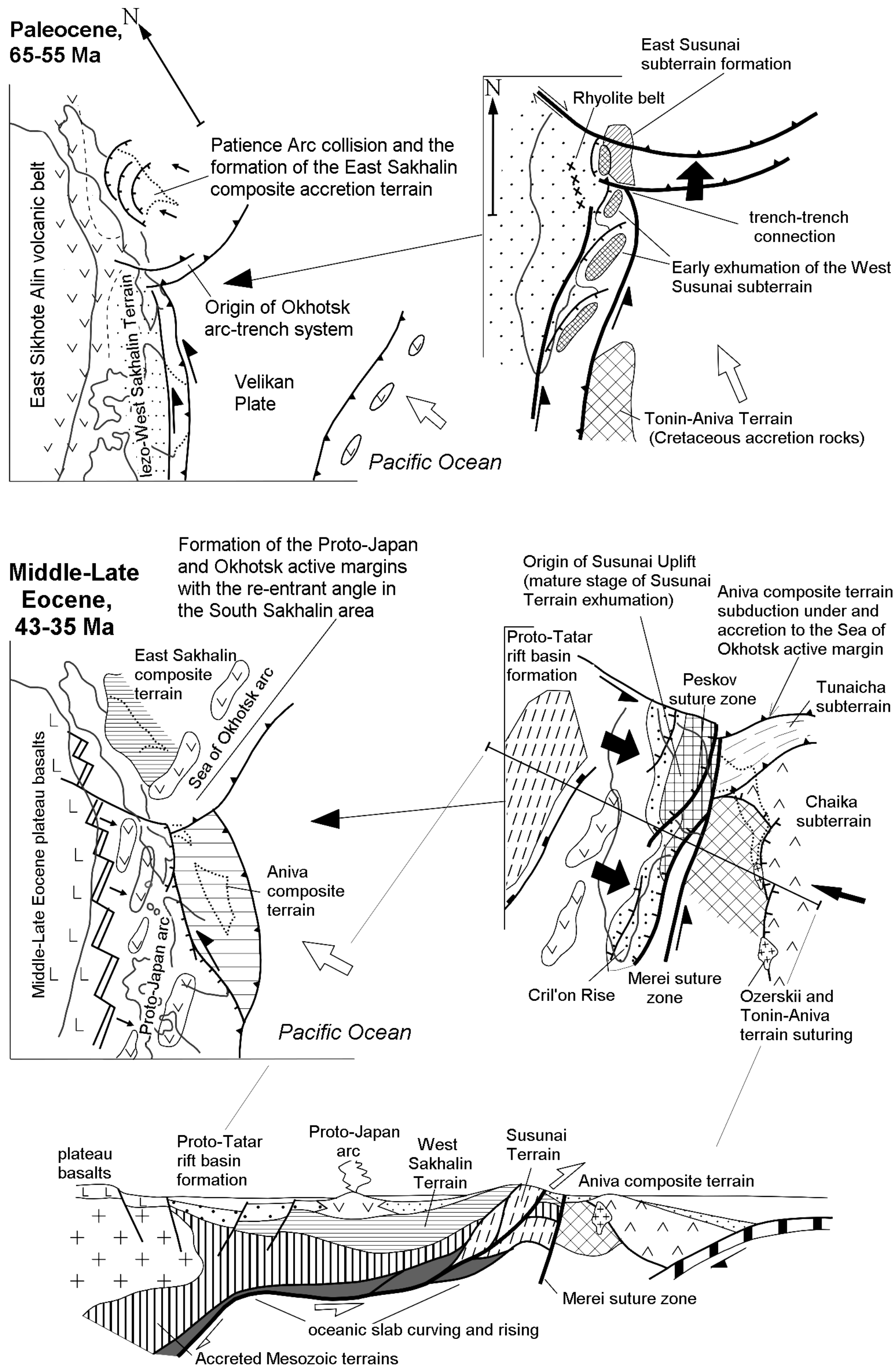

Figure 14. South Sakhalin Paleocene-Eocene tectonic reconstructions and a model for the Susunai Terrain tectonic floating up.

ison of the specific structural and petrological features of the proto-Japan Arc and the adjacent structural features suggests their evolution as a result of the break or the abrupt curvature of the previously subducted slab of the oceanic crust and its subsequent "floating up" or rolling away toward the trench. These movements "shaped" the late metamorphic structural features in the Susunai Terrain and controlled the final exhumation of the metamorphic rocks.

[136] The Okhotsk arc-trench system was being formed at that time in the area of the Sea of Okhotsk continental margin. Its frontal elements are preserved in the East Susunai and Tunaicha subterrains. Its Eocene-Oligocene volcanic rocks (46–26 Ma), marking the zone of the suprasubduction igneous activity, have been found in the southern part of Central Sakhalin and in its sea-floor extension, recorded in the Academy of Science Rise [*Emel'yanova,* 2003].

Conclusion

[137] The formation of the heterogeneous accretion-type structure of Southeast Sakhalin continued from the Aptian to the Middle-Late Eocene during the repeated rearrangements of its convergent boundary, caused either by the accretion of or by the collision with the continental margin of the intraoceanic terrains, or by their subsequent movements together with the continental-margin accreted rock complexes. During its Cretaceous-Early Paleogene evolution the East Asia continental margin varied in its geodynamic conditions from those of the active margin of the West Pacific type, combined with the transform-type margin of the Californian type [after *Khanchuk*, 1993] during the Early Cretaceous, comparable with the environment similar to that of modern Polynesia (with its island arcs displaced to the ocean and vast marginal seas) during the Campanian-Eocene.

[138] The evolution of the continental margin had a cyclic character and was interrupted by the epochs of global and regional structural rearrangements, recorded at five age levels: (1) the period from the end of the Neocomian to the beginning of the Aptian was marked by the displacement of the continent-ocean boundary from the eastern Sikhote Alin area, and by the formation at the oceanic basement of the accretion-type structural features of the Hokkaido-Sakhalin fold system, which developed under the conditions of the oblique, left-lateral subduction; (2) the boundary between the Early and Late Cretaceous was marked by the crowding of the continental margin, island arc, and oceanic terrains in the Hokkaido-South Sakhalin-Tatar Strait area as a result of the large-scale strike-slip movements along the continental margin; (3) the formation during the Early Campanian of the system of ensimatic island arcs, separated from the continent by epioceanic marginal seas; (4) the formation during the Paleocene of a collision-type structural feature in the eastern areas of Central Sakhalin and the origin, at its southern flank, of the sublatitudinal structural features of the Okhotsk active continental margin, which in South Sakhalin, were connected, in mullion manner, with the Cretaceous structural features of the East Asian margin; (5) the Middle Eocene witnessed the collision of the Tokoro ensimatic arc with the proto-Japan continental margin, the formation of the Aniva composite terrain, and the migration of the convergent boundary over a large distance farther east and southeast. Each of these epochs concluded the respective period of the growth of the continental margin and caused a change in the mechanism of its accretion.

[139] The characteristic feature of the Cretaceous-Paleogene evolution of the East-Asian continental margin was the predominant growth of its Sea of Okhotsk segment in the southern direction (from the Okhotsk-Chukotka volcanic belt to South Sakhalin and to the Academy of Science High) by more than 1000 km. At the same time, the Sea of Japan sector of the continental margin grew in the eastern direction (from the Jurassic-Early Cretaceous accretion-type structural features of the Sikhote Alin Range to the accretion-type structural features of East Sakhalin and Hokkaido) by merely a few hundred kilometers, taking into account the closed Sea of Japan. These specific features can be explained by the different mechanisms of the accretion, namely, by the alternation of the processes of the primarily orthogonal subduction and collision of the intraoceanic terrains in the Sea of Okhotsk sector, and by the left-lateral oblique subduction and the transform-type movements of the polygeodynamic terrains along the continental margin.

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