(2) main regularities in distribution of paleogeographic settings in the continent/ocean transition zone.

Most of cited publications that were used when compiling lithologic–paleogeographic maps were listed in previous works (Zharkov *et al.*, 1995, 1997). The reference list of this paper includes only newer, additional works.

MAIN FEATURES OF PALEOGEOGRAPHY DURING THE CONIACIAN–MAASTRICHTIAN

The second half of the Late Cretaceous represents a transition period from the Mesozoic to Cenozoic planetary paleogeographic situation. It was a period when the Earth's surface was still divided into oceanic hemisphere occupied by the Pacific and continental hemisphere comprising all the continental blocks: Laurasia, South America, Africa, Hindustan, and Eastern Gondwana, which included Antarctica, Australia, and New Zealand at the beginning of the second half of the Late Cretaceous. The global paleotectonic and paleogeographic asymmetry of the Earth was responsible for peculiarities in many paleogeographic reorganizations that occurred in oceanic and continental hemispheres throughout this period.

The continental hemisphere was characterized by a mosaic distribution of continents separated by oceans of sublatitudinal and submeridional strikes. This hemisphere retained a paleogeographic asymmetry typical of the entire Cretaceous. The asymmetry was reflected in the opposite position of the huge Laurasian continent, located mainly in temperate and high latitudes of the northern hemisphere, on the one hand, and in the isolated Gondwanan continental blocks of South America, Africa, Hindustan, and Eastern Gondwana, situated mostly in the southern hemisphere, on the other. The Tethys ocean stretching in the sublatitudinal direction along the northern tropical and subtropical belts was a structure separating these asymmetric paleogeographic segments. The processes that controlled initiation and opening of new oceans and interrelated breakup events and movements of continental blocks continued in the southern hemisphere.

In the southern hemisphere, the progressing extension of the South Atlantic and South Ocean occurrred throughout the entire second half of the Late Cretaceous (Sclater *et al.*, 1977; Rabinowitz and La Brecque, 1979; Barron *et al.*, 1981; Ziegler *et al.*, 1982; Emery and Uchupi, 1984; Zonenshain *et al.*, 1984, 1990; Krasheninnikov and Basov, 1985; Barron, 1987; Scotese *et al.*, 1988; Sager and Scotese, 1989; Scotese, 1991; Zonenshain and Kaz'min, 1992; Khain and Balukhovskii, 1993; Golonka *et al.*, 1995). The width of the South Atlantic increased by approximately 2000 km (from 3500 km in the Coniacian to more than 5500 km in the Maastrichtian) in extreme southern areas and by 1200 km in its northern areas, where it became 2500 km wide in the Maastrichtian. Simultaneous extension of the North Atlantic improved connections between South and North Atlantic and resulted in formation of the intergrated Atlantic Ocean of the submeridional strike. The South Ocean stretched near 60°S from the South America on the west to Australia on the east occupying areas between Antarctica, Africa, and Hindustan. Its widths varied from 2000–2500 km to 3000–3500 km. In its central part, the South Ocean was connected with the Eastern Tethys via the deep Mozambique–Somalia seaway, and, in the eastern part, it freely opened into the Himalayan–Indonesian Tethys and Pacific.

In the Campanian, Australia was separated from Antarctica due to southeastern and eastern progradation of the spreading axis, which resulted in the formation of another deep seaway between the Pacific and South oceans (Veevers, 1984; Patriat and Segoufin, 1988; Scotese *et al.*, 1987, 1988; Scotese, 1991). The separation of New Zealand from Australia and Madagascar from Hindustan also occurred at that time.

While the continental breakup with initiation and opening of deep oceanic basins proceeded in the southern segment of the continental hemisphere, radial subsidence continued in the northern segment of Laurasia, where simultaneous formation of epicontinental seas of submeridional or similar strike divided the Laurasia continent into autonomous land blocks. In the west, the Western Interior sea basin connecting the Caribbean part of the Tethys with the Amerasian ocean continued to exist throughout the entire time period in question; it separated the land area of the Rocky Mountains, stretching along the western periphery of the North American margin of Laurasia, from the Canadian lowland. In interior areas of Laurasia between the Canadian lowland and Greenland and also between Greenland and Baltia, the Labrador–Baffin and Norwegian seaways connected the North Atlantic with the Amerasian ocean. In the east, the Laurasia continent always hosted the inland West Siberian sea, intermittently connected, via the Turgai seaway, with the Tethys-related epicontinental seas. Eastern Laurasia was occupied by a spacious land, Angarida, whose eastern and southern margins hosted mountain systems, whereas northern and western areas were occupied by lowlands. From the Coniacian to Maastrichtian ages, huge northern and southern areas of Laurasia were occupied, along with above-mentioned seaways, by spacious epicontinental seas. Its southern margins were covered by the interconnected North, Polish–Lithuanian, Central European, East European, Turan, and Afghan–Pakistan seas, whereas the Sverdrup, North Greenland, Barents, Kara, and other shelf and epicontinental seas existed along the northern margin. More than 40% of Laurasia was covered by seas. In the paleogeographic respect, this feature substantially distinguished this continent from all southern continental blocks of the Gondwana group, a significant part of which constantly represented land areas. Only the Trans-Sahara seaway similar, in some features, to Laurasian seaways existed in North Africa.

However, the Trans-Sahara seaway was located in the tropical and equatorial zones, and this resulted in a combination of evaporitic and normal marine sedimentation. During the second half of the Late Cretaceous, paleogeographic patterns of the continental hemisphere resembled the situation that appeared beginning in the mid-Cretaceous and was related to the asymmetric position of epicontinental and shelf seas in Laurasia, on the one hand, and in southern continents, on the other.

According to recent paleogeographic reconstructions (Philip *et al.*, 1993a, 1993b; Camoin *et al.*, 1993a, 1993b), the Late Cretaceous Tethys was subdivided into five main region replacing one another in the west–east direction: (1) the Caribbean, located between two open oceans, the Pacific on the west and the Atlantic Tethys on the east; (2) the Atlantic, which represented a symmetric oceanic basin with the mid-ocean ridge; (3) the Mediterranean, consisting of the mosaic system of shallow carbonate platforms separated by deep sea basins often floored by oceanic crust; (4) the Eastern Tethys, comprising oceanic areas between Eurasia and the Arabian margin of Africa; and (5) the Himalayan–Indonesian Tethys, which covered areas between the Asian part of Laurasia on the north and Hindustan and Australia on the south. At the same time, one should remember that a series of important paleogeographic events, which resulted in principal changes in configuration and extension of the Cretaceous Tethys, occurred during the second half of the Late Cretaceous, particularly in the Campanian. They were related to the transformation of Atlantic basins into the integrated ocean extending in submeridional direction. In every paleotectonic and paleogeographic respect, the North Atlantic was united with the South Atlantic into a single planetary-scale ocean at that time. The North Atlantic up to 5000 km wide represented an oceanographic barrier that isolated the Caribbean realm from the Mediterranean province (Klitgord and Schouten, 1986; Camoin *et al.*, 1993b). That was the reason why the North Atlantic was no more a constituent of the Tethys, but became an element of the Atlantic Ocean, which divided the sublatitudinal Cretaceous Tethys into two isolated and autonomous parts unequal in size: the western Caribbean part and the eastern one comprising the Mediterranean, Eastern, and Himalayan–Indonesian Tethys. Beginning from the terminal Cretaceous, only this eastern part should be considered as the Tethys proper.

Despite the Tethys fragmentation, the more-or-less free water exchange with the Pacific continued to exist in the east and west, and gave rise to circum-global Western currents in tropical latitudes of the northern hemisphere (Luyendyck *et al.*, 1972; Gordon, 1973; Berggren and Hollister, 1974; Camoin, 1993). It is assumed that circular clockwise surficial currents were quite significant in the North Atlantic, Mediterranean, and Eastern Tethys (Barron and Peterson, 1989; Tucholke and McCoy, 1986). It should be noted that the free influx of warm tradewind currents from the Pacific into the South Ocean and further into the South Atlantic occurred via the wide seaway between Hindustan and Australia and through the Mozambique–Somalia strait. As for the surficial water exchange with the Amerasian ocean, it is assumed that respective currents passed through the Labrador–Baffin and, probably, Norwegian seaways (Berggren and Olson, 1986; Gradstein *et al.*, 1991).

The Pacific hemisphere of the Earth represented a deep oceanic basin, which had free exchange with the Tethys and South Ocean via wide seaways in the west and restricted connection with the Caribbean province and Atlantic Ocean in the east. In the tropical latitudes, the circular western current passed through the entire Pacific. It is presumed that spacious anticyclonic circulation existed to the south and north of the tropical currents.

The major central part of the Pacific Ocean bottom was occupied by the Pacific plate, which gradually extended owing to the spreading. The Kula (on the north), Farallon (on the east), and Phoenix (on the southeast) plates bounding the Pacific were reduced in area as a result of subduction. According to paleotectonic reconstructions (Scotese *et al.*, 1988; Kononov, 1989; Pushcharovskii and Melankholina, 1992) that were modified by M.V. Kononov in our maps, the sublatitudinal part of the spreading ridge between the Pacific and Kula plates migrated northward from 10°–20°N in the Coniacian to 25°–35°N in the Maastrichtian. The submeridional part of the ridge migrated eastward nearing the position of the modern East Pacific Rise at the end of the Cretaceous. Correspondingly, intraplate morphostructures (undersea mountains and rises, deep-sea basins) located on the Pacific plate also migrated together with their facial complexes in the course of spreading. Simultaneously, lithospheric plates with related bottom morphostructures experienced the isostatic subsidence.

During the entire time period in question, the central part of the Pacific (including the entire Pacific plate) reprsented a spacious pelagic area, which was characterized by extremely low terrigenous influx from continents, low bioproductivity of surficial waters, and active aeration of bottom waters. Under such conditions, red (oxidized) pelagic sediments were slowly accumulated: calcareous nannoplanktonic–foraminiferal oozes in the undersea rises above the carbonate compensation depth (CCD) and siliceous (radiolarian) to clayey deposits (pelagic brown clays) in abyssal plains below the CCD (Murdmaa, 1987). The CCD level, i.e., the boundary between the carbonate and carbonate-free facis, which was highest in the mid-Cretaceous, slightly lowered by the end of this period. Correspondingly, carbonate sediments prograded to greater depths.

From the Coniacian to Maastrichtian, abyssal areas (below the CCD, i.e., 4–5 km deep) of the Pacific and, probably, neighboring oceanic plates gradually deepened and widened that extended areas of carbonate-free clayey and siliceous–clayey sedimentation. In addition to Central Basin, the South Basin was formed, and brown pelagic clays, the indicators of low bioproductivity zones ("oceanic deserts"), were deposited here. In the Central Basin, which gradually moved northward, two facial complexes, pelagic clayey and clayey–radiolarian, were accumulated. Some data indicate that siliceous sediments were mostly confined to the southwestern equatorial part of the basin, where a sublatitudinal belt of silica accumulation, probably related to the above-mentioned circum-tropical currents, was formed before the Maastrichtian.

In the intraplate rises and mid-oceanic ridges (above the CCD), pelagic sediments accumulated.

The northward migration of the Mid-Pacific Mountains was in progress, and they reached the equatorial zone and subsided below sea level to depths of several hundreds meters to be transformed into guyots. Bottom currents washed the guyot tops and created environments of nondeposition or extremely slow pelagic sedimentation (strongly condensed sequences) and phosphate and ferromanganese hardgrounds (Murdmaa *et al.,* 1995). The volcanic activity continued south of the Mid-Pacific Mountains, in the Darwin swell, and resulted in the formation of new seamounts of the modern archipelago of the Marshall Islands.

MAIN REGULARITIES IN DISTRIBUTON OF PALEOGEOGRAPHIC SETTINGS IN THE CONTINENT–OCEAN TRANSITION ZONES

At the end of the Cretaceous (88–65 Ma), paleogeographic settings (and corresponding lithologic complexes) of the continent–ocean transition zone were primarily controlled by the interaction between oceanic and continental plates. For instance, relatively slow widening of the Atlantic (Zonenshain and Kuz'min, 1992) favored the quiet and stable tectonic regime in its continental framing and the formation of a relatively narrow band of shallow-water terrigenous deposits along the periphery, which graded seaward into hemipelagic clayey and carbonate–clayey facies. By contrast, in the Pacific and Tethys, converging oceanic and continental plates produced quite different landscapes of the continent–ocean transition zone.

Continental marginal zones of the Pacific had a complicated structure. In the second half of the Late Cretaceous, the northeastern peripheral part of the Pacific was characterized by a combination of strike-slip movements and subduction of oceanic plates under the North American continent (Csejtey *et al.,* 1982; Monger *et al.,* 1982; Coney, 1987; Frazier and Schwimmer, 1987; Plafker *et al.,* 1989; Wallace *et al.,* 1989; Haugerud *et al.,* 1991; Livaccari, 1991; Underschultz and Erdmer, 1991; Ingram and Hutton, 1994). A series of narrow undersea mountain chains with relatively shallow-water terrigenous sedimentation was formed along the peripheral zone of the ocean during the Coniacian–Santonian (Figs. 1, 2). They represented terranes, including the most spacious Talkeetna terrane. The mountain chains were bounded by troughs with turbidite accumulation in both continental and oceanic sides. Eastern structures of the last type are usually considered as marginal troughs relative to the North American continent. In the Wrangellia and Talkeetna mountains, the thickness of the Coniacian–Campanian flysch is up to 1000 m (Plafker *et al.,* 1989). Similar flysch deposits are also known from southern areas, e.g., from western California (Khain and Balukhovskii, 1993). Subsequent strike-slip movements resulted in the migration of the Talkeetna terrane and corresponding mountain chain (Figs. 1–3). In the Campanian, the latter joined the western margin of North America, which resulted in the formation of thick (800–2000 m) silty–sandy coal-bearing deposits intercalated with conglomerate beds near the continent edge.

The western margin of the North American continent was characterized in the second half of the Late Cretaceous by a permanent uprising and formation of the Cordillera orogenic system. Two main factors were responsible for this last event: the rise of the continental margin and intense eastward horizontal movements of rock masses (the Laramian orogeny). These movements resulted, first, in the thrust–nappe structure of mountains and, second, in asymmetric patterns of distribution of continental and coastal basins along the eastern Cordillera piedmont. Orogeny was most intense during the Campanian–Maastrichtian, as it is reflected in deposits of synorogenic basins adjoining the Cordilleras in the east (Coney, 1987). The area of terrestrial sedimentation between the Cordilleras and Western Interior seaway was of a complex structure (Dickinson *et al.,* 1988). Three zones are distinguished in the Rocky Mountains located within this area. The western zone that directly joined the Cordilleras included the so-called barrier basins, where sedimentation was most intense and long-lasting. The intermediate zone consisted of alternating ridges and depressions. The third eastern zone included peripheral basins. In the Coniacian–Maastrichtian, small but numerous basins of terrestrial sedimentation existed along strike-slip faults in the Alaskan Cordilleras (Fisher *et al.,* 1982). Thick Upper Cretaceous terrigenous sequences filling these basins incorporated coal-bearing members, whose abundance was particularly significant at the Cretaceous–Paleogene boundary (Fig. 4).

Inasmuch as the intensity of subduction sharply decreased in the interval from the Turonian to Campanian and strike-slip movements became prevalent, the type of magmatism also changed in the west of North America. As shown earlier (Zharkov *et al.,* 1995; Filatova, 1996), in the Albian–Cenomanian, this area included the lateral structural succession typical of the active continental margin: a deep-sea trench marking the subduction zone; a fore-arc trough with turbidite accumulation; and a plutonic belt of the continental margin. The second half of the Late Cretaceous was marked by the gradual cessation of activity in this belt (Figs. 1, 2). Judging from discrete patterns of magmatic rock ages (Evenchick, 1991), the magmatic activity was pulsating against the background of constant continental margin rising and displacement of rock sequences eastward along overthrust surfaces and parallel to the continent–ocean boundary along strike-slip faults.

The formation of the Laramide orogenic belt and corresponding Cordillera system in the Campanian–Maastrichtian changed the type of magmatism and resulted in the appearance of muscovite granitoids in California (Karlstrom *et al.,* 1993). Further development of the Laramian orogeny was accompanied by basaltic volcanism in the strike-slip zones (Dickinson *et al.,* 1988; Plafker *et al.,* 1989) (Fig. 4).

Three segments (northern, central, and southern), which show certain differences in their evolution, can be distinguished in the southeastern peripheral area of the Pacific and in the west of South America during the Late Cretaceous (Herve *et al.,* 1987; and others).

The most peculiar history was that of the Venezuela–Ecuador segment, where the synchronous formation of perioceanic island-arc and continental–marginal volcanic belts occurred in the Late Cretaceous (McCourt *et al.,* 1984; Megard, 1987; Aguirre *et al.,* 1992; Van Thrournout *et al.,* 1993; Desmet, 1994; and others). During the Laramian orogeny, which commenced in the Campanian and continued till the early Paleogene (Megard, 1987), the island-arc belt was obducted onto the continent. This resulted in crustal thickening of the northwestern margin of the South American continent and the formation of the Laramian thrust–nappe orogenic belt comprising the huge structure of the Andes. The eastern front of the continental margin was occupied by the synorogenic basin with terrigenous sedimentation. During the Campanian–Maastrichtian, the northwestern margin of the South American continent was involved in the marine sedimentation cycle lacking any significant volcanism (Figs. 3, 4).

The central segment of western South America hosted the Araukan continental–marginal volcanic belt (Lomize, 1983), whose formation commenced after the Peruvian (Austrian) orogeny. Continental volcanic deposits in the central segment of this belt are up to 4000–6000 m thick, narrowing northward down to 1000 m and less. During the Early–Late Mesozoic period, the axis of continental–marginal volcanic belt migrated eastward (James, 1971; Coira *et al.,* 1982; Hevre *et al.,* 1987) in response, as usually explained, to the gradual flattening of the subduction zone plane. Westward of the central segment of the Andes, the Coastal Range zone of uplifts and the fore-arc trough with terrigenous–volcanogenic deposits were located. In the east, the volcanic belt was in contact of a zone of uplifts, which, in turn, joined, further to the east, the submeridional chain of back-arc basins with terrestrial deposits and volcanics. The latter structures were represented by the Subandean and Andean basins of Bolivia and Peru, as well as by the Chacoparana and Salta basins of Argentine (Grier *et al.,* 1991).

The continental–marginal volcanic belt of the central segment evolved through the entire Late Cretaceous. In the Coniacian–Santonian, it extended into the southern segment of South America and further to the south along the Antarctica margin. The southern segment, to the east of the volcanic belt, hosted the mountain chain that was the eastern boundary of the Magellan marginal basin. This zoning in the southern segment of western South America, which appeared in the late Albian, existed till the Campanian. The Laramian orogeny resulted in the general rise of the continental margin, in the cessation of volcanic activity and in the eastward migration of the Magellan basin (Wilson, 1991).

In general, the Late Cretaceous was marked by the development of two types of basins in the western margin of South America behind the volcanic belt: (1) the basins genetically related to mountainous chains migrating eastward and generally marked by terrigenous sedimentation; and (2) back-arc (relative to the volcanic belt) basins of rift origin with alkaline and bimodal magmatism. Basins of the first type were widespread along the entire western margin of South America, whereas local rift basins were mainly characteristic of the central segment of the South American continent during the Late Cretaceous.

The northwestern and western peripheral areas of the Pacific along the Asian continental margin hosted a chain of active volcanoes of an extended continental–marginal volcanic belt with several diachronous segments: Okhotsk-Chukotka, Sikhote-Alin, and Cathasian. The last one (*Atlas* ..., 1985) included volcanic structures of the maritime part of China, together with those of southern Korea and northern Japan, now separated by the Sea of Japan (Filatova, 1990; Chang, 1995).

The continental–marginal volcanic belt was associated with a deep-sea trench (Figs. 1–3). The inner, continental slope of the trench was of a zoned structure. The zone near the continent was occupied by a shallow sea with abundant islands and island chains. During the early epochs of the Late Cretaceous, the sea accumulated flyschoid sequences, which later were replaced by coarse olistostrome–molasse complexes (Zinkevich, 1981; Filatova, 1995, 1996). Futher seaward, thick turbidite sequences were formed on the deeper slope of the trench. They are exemplified by the Upper Cretaceous sequences of western Sakhalin (Rozhdestvenskii, 1987; Zonenshain *et al.,* 1990) and by deposits of the Simanto and Hidaka zones of Japan (Taira, 1985; Kimura *et al.,* 1990). An ensemble of volcanic island arcs and marginal seas was located in the perioceanic zone (Bogdanov and Filatova, 1988; Zonenshain *et al.,* 1990; Sokolov, 1992); one of its segments was represented by the Irunei island arc. This structural ensemble constantly migrated landward and joined the continent in the Maastrichtian (Mitrofanov, 1977; Alekseev, 1979; Bogdanov *et al.*, 1982; Filatova, 1988; Bogdanov and Til'man, 1992; Parfenov *et al.*, 1993). Fragments of its island arcs are preserved in the form of tectonic nappes in Kamchatka and eastern Sakhalin (Pushcharovskii *et al.*, 1983; Pushcharovskii and Melankholina, 1992; and others), as well as in Hokkaido (Taira, 1985, Kimura and Kensaku, 1986; Maruyama and Seno, 1986; and others).

The considered events occurred during the Laramian orogeny, which was characterized by the general rise of the Asian margin (Zinkevich, 1981; Ruzhentsev *et al.*, 1982; and others), where isolated basins accumulated coal-bearing molasses. This orogeny caused cessation of volcanism in some segments of the continental–marginal volcanic belt (for instance, in the Okhotsk-Chukchi belt), although other segments continued to be active till the beginning of the Paleogene.

Reorganization of oceanic plates in the northwestern and western parts of the Pacific at 85–74 Ma was noted by many researchers (Kimura *et al.*, 1985, Maruyama and Seno, 1986; Jolivet *et al.*, 1988; Zonenshain *et al.*, 1990; Khain and Balukhovskii, 1993; Parfenov *et al.*, 1993; Okada and Sakai, 1993; and others). In the Campanian–Maastrichtian (Figs. 3, 4), the Sea of Okhotsk represented an uplifted area occupied by a shallow sea with island chains and bounded by the volcanic belt in the south. (Maruyama and Seno, 1986). In the continental margin, this time period was marked by the formation of pull-apart basins, which accumulated terrigenous coal-bearing deposits. Such basins, which are often of a half-graben form, were long known in China and recently were also discovered in Japan (Okada and Sakai, 1993). Movements along faults that confine these basins resulted in the formation of relatively small, but abundant basaltic plateaus (Filatova, 1988).

As for the southwestern periphery of the Pacific, important events occurred there in the terminal Late Cretaceous. Whereas the Coniacian–Santonian time was characterized by the formation of the volcanic belt on the Australian–Antarctic margin (Howell, 1980), the Campanian was marked by the opening of the oceanic basin between Australia and Antarctica and by the formation of the Tasman Sea bounded by the volcanic island arc on the Pacific side (Veevers, 1984; Scotese *et al.*, 1988) (Figs. 3, 4).

Drastic changes in the plate movement occurred in the Tethyan region during the Late Cretaceous (Dercourt *et al.*, 1985, 1993; Zonenshain *et al.*, 1987). The Eurasian and African–Arabian blocks have exchanged their previous southward migration for northward movements. The Bay of Biscay opened, and the Apulian block was separated from Africa. The oceanic plate of the Eastern Carpathian and Balkan basins subducted southward to give rise to the formation of volcanic belts. The Late Cretaceous subduction of the oceanic crust under Eurasia resulted in formation of the major South Eurasia volcanic belt along the northern margin of the Tethys (Kaz'min *et al.*, 1987). The western extremity of the latter is recognized in the Bulgarian Highland. Eastward, the belt is traced in the Western Pontids and further to the south, in the echelon-like structures of the Eskisehir Mountains, where volcanism commenced in the Turonian (Sengor and Yilmaz, 1981). In the Transcaucasian segment of the volcanic belt, a thick volcanic sequence was formed in the latest Albian–early Campanian, whereas Campanian–Maastrichtian volcanics are much less widespread here. In general, two stages of the volcanic activity can be distinguished in the Pontian–Transcaucasian segments of the belt: the Albian–early Campanian and late Campanian–Maastrichtian. The main volume of volcanics was accumulated during the first stage. The volcanics are referred to calc-alkaline–alkalic and, less commonly, to tholeiitic magmatic series (Lordkipanidze, 1980). The second stage was marked by basaltic eruptions of subalkalic composition (Kaz'min *et al.*, 1987); this allows an assumption that the last stage was not subduction-related and induced after cessation of subduction during the Laramian orogeny.

The eastern continuation of the Late Cretaceous belt is traced in the northwestern Elburs region. Further to the east, calc-alkaline volcanics are recorded in tectonic fragments of the Mekran ophiolites between the Dgaz-Murian basin in the north and Badgan-Dur-Kan zone in the south. Still further to the east, volcanics of the Late Cretaceous age occur in Afghanistan.

At the beginning of the Coniacian Age, the Cyprus ensimatic island arc appeared in the southern part of the Tethys (Dercourt *et al.*, 1985; Knipper, 1985; Zonenshain *et al.*, 1987). In the Indian Ocean, the Kohistan (Bard, 1983), or Dras (Dietrich *et al.*, 1983) intraoceanic island arc with associated deep-sea trench in the south was formed in the time period of 90–75 Ma (Figs. 1, 2). Tholeiites erupted in the arc (Dietrich *et al.*, 1983) and volcanogenic–terrigenous deposits were accumulated in the fore-arc basin (Robertson and Degnan, 1994).

In the terminal Cretaceous time and at the Cretaceous–Paleogene boundary, the Tethyan region experienced intense orogeny (synchronous with the Laramian one), which resulted in the closing of the southern part of the Eastern Tethys and in a collision between Africa and Eurasia (Dercourt, 1985, 1993; Zonenshain *et al.*, 1987; Knipper and Sharas'kin, 1995). Sedimentary sequences of the Alpine and Carpathian–Pannonian basins were deformed. The Late Cretaceous volcanism in the Carpathians ceased. At the end of the Late Cretaceous (approximately 80 Ma), the African–Arabian margin was subducted under the Cyprus arc and ophiolite nappes were obducted onto the continental margin. The Campanian was marked by a collision of the Kohistan island arc with the northern margin of India

(Bard, 1983; Zonenshain *et al.*, 1984). In the modern structure, tectonic nappes of island-arc volcanics and fore-arc basinal complexes occur in the Ind-Zangpo collision suture, which was formed in the Eocene (Allegre *et al.*, 1984).

The Late Cretaceous collision was accompanied by the gradual degeneration of the South Asian volcanic belt, whose activity had slowed down since the Campanian and almost completely stopped in the Maastrichtian (Fig. 3). Tectonic compression and obduction along the southern boundary of the Tethys were accompanied by substantial extension of back-arc basins in the Eurasian continental margin. These basins formed a system of Paratethyan seas 900 km wide and stretching from the Balkans to eastern Iran. Two (Western and Eastern) Black Sea basins opened, and the Central Iran oceanic basin between Lut, Sanandadg-Sirdgan, and Elburs structures increased in dimensions at that time.

CONCLUSION

The following main peculiarities of paleogeographic reorganizations that occurred during the second half of the Late Cretaceous can be inferred from the analysis of lithologic–paleogeographic maps compiled for the Coniacian, Santonian, Campanian, and Maastrichtian ages.

(1) Many features of paleogeographic reorganizations were related to the global paleotectonic and paleogeographic asymmetry of the Earth. The asymmetry was primarily reflected in the contraposition of oceanic and continental hemispheres, as well as in the contraposition of Laurasian continent in the north and isolated Gondwanan continental blocks of the south in the last hemisphere. The southern segment of the continental hemisphere was characterized by the development of oceans, which appeared in earlier epochs, and by origin of new oceanic basins (South Atlantic, South Ocean, Mozambique–Somalia and others) associated with break-up and migration of continental blocks (South America, Africa, Hindustan, Antarctica, New Zealand). Simultaneously, radial subsidence and formation of sublatitudinal epicontinental seas occurred in the northern segment of Laurasia. These asymmetric segments of the continental hemisphere were divided by the sublatitudinal Tethys Ocean. After integration of Atlantic basins into a single global structure at the end of the Late Cretaceous, the Tethys was diveded into two independent parts: the Caribbean basin in the west and Late Cretaceous Tethys proper in the east, where it included the Mediterranean, Eastern, and Himalayan–Indonesian parts. The oceanic hemisphere was occupied by the Pacific. The Pacific plate of its central part was gradually extending due to spreading, whereas the Kula, Farallon, and Phoenix plates were shrinking owing to subduction. During the entire second half of the Late Cretaceous, the central part of the Pacific represented a deep pelagic province with gradually extending and deepening abyssal plains.

(2) The paleogeographic situation in the ocean–continent transition zones was controlled by the interaction between oceanic and continental plates. Most significant paleogeographic reorganizations occurred in the peripheral zones of the Pacific and Tethys and also in continents around them.

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REFERENCES

Alekseev, E.S., Characteristic Features of the Evolution and Structure of the Koryak Upland, *Geotektonika*, 1979, no. 1, pp. 85–95.

Barron, E.J. and Peterson, W.H., Model Simulation of the Cretaceous Ocean Circulation, *Science,* 1989, vol. 244, pp. 684–686.

Berggren, W.A. and Olson, R.K., North Atlantic Mesozoic and Cenozoic Paleobiogeography, in *The Geology of North America: The Western North Atlantic Region,* Vogt, P.R. and Tucholke, B.E., Eds., Boulder: Geol. Soc. Am., 1986, pp. 231–255.

Bogdanov, N.A., Chekhovich, V.D., Sukhov, A.N., and Vishnevskaya, V.S., Tectonics of the Olyutorskaya Zone, in *Ocherki tektoniki Koryakskogo nagor'ya* (Tectonic Reviews on the Koryak Upland), Moscow: Nauka, 1982, pp. 189–217.

Camoin, G., Bellion, Y., Benkhelil, J., *et al.*, Late Maastrichtian Palaeoenvironments (69.5–65 Ma), in *Atlas of Tethys Palaeoenvironmental Maps,* Dercourt, J., Ricou, L.E., and Vrielynck, B., Eds., Paris: Rueil-Malmaison, 1993a.

Camoin, G., Bellion, Y., Dercourt, J., *et al.*, Late Maastrchtian (69.5–65 Ma), in *Atlas of Tethys Palaeoenvironmental Maps: Explanatory Notes,* Dercourt, J., Ricou, L.E., and Vrielynck, B., Eds., Paris: Gauthier-Villars, 1993b, pp. 179–196.

Chang, K.H., Aspects of Geologic History of Korea, *J. Geol. Soc. Korea,* 1995, no. 31, pp. 72–90.

Desmet, A., Le magmatism ophiolitique de la Quebrada San Juan et les basalts de la serie Pinon (Equateur), *Ophioliti,* 1994, vol. 19, no. 1, pp. 115–156.

Filatova, N.I., *Periokeanicheskie vulkanogennye poyasa* (Perioceanic Volcanic Belts), Moscow: Nedra, 1988.

Filatova, N.I., The Mesozoic Evolution of the Korea–Japan Region, *Geotektonika,* 1990, no. 5, pp. 112–124.

Filatova, N.I., Evolution of Cretaceous Environments in the Northeast of the Asian Continent, *Stratigr. Geol. Korrelyatsiya,* 1995, vol. 3, no. 3, pp. 64–77.

Filatova, N.I., The Middle Cretaceous Evolution of Active Margins, *Geotektonika,* 1996, no. 2, pp. 74–98.

Fisher, M.A., Patton, W.W., Jr., and Holmes, M.Z., Geology of Norton Basin and Continental Shelf beneath Northwestern Bering Sea, Alaska, *Bull. Am. Assoc. Petrol. Geol.,* 1982, vol. 66, no. 3, pp. 255–285.

Golonka, J., Ross, M.I., and Scotese, C.R., Phanerozoic Paleogeographic and Paleoclimatic Modeling Maps, *Mem. Can. Soc. Petrol. Geol.,* 1995, no. 17, pp. 1–47.

Gradstein, F.M. and Rad, U. von, Stratigraphic Evolution of Mesozoic Continental Margin and Oceanic Sequences: North Australia and Northern Himalaya, *Marine Geol.*, 1991, vol. 102, pp. 131–173.

Haugerud, R.A., Van Der Heyden, P., Tabor, R.W., *et al.*, Late Cretaceous and Early Tertiary Plutonism and Deformation in the Skagit Gneiss Complex, North Cascade Range, Washington and British Columbia, *Bull. Geol. Soc. Am.*, 1991, vol. 103, no. 10, pp. 1297–1307.

Herve, F., Godoy, E., Parada, M., *et al.*, General View on the Chilean–Argentine Andes, with Emphasis on Their Early History, *Circum-Pacific Orogenic Belts and Evolution of the Pacific Oceanic Basin*, Monger, J.W.H. and Francheteau, J., Eds., Boulder: Geol. Soc. Am., 1987, pp. 97–111.

Ingram, G.M. and Hutton, D.H.W., The Great Tonalite Sill: Emplacement into a Conractional Shear Zone and Implications for Late Cretaceous to Early Eocene Tectonics in Southern Alaska and British Columbia, *Bull. Geol. Soc. Am.*, 1994, vol. 6, no. 5, pp. 715–728.

James, D.E., Plate Tectonic Model for the Evolution of the Central Andes, *Bull. Geol. Soc. Am.*, 1974, vol. 82, no. 12, pp. 3325–3346.

Jolivet, L., Cadet, J.P., and Lalevee, F., Mesozoic Evolution of Northeast Asia and Collison of the Okhotsk Microcontinent, *Tectonophysics*, 1988, vol. 149, no. 3/4, pp. 89–109.

Karistrom, K.E., Miller, C.F., Kingsbury, J.A., and Wooden, J.L., Pluton Emplacement in Active Ductile Thrust Zone. Plute Mountains, South California: Interaction between Deformational and Solidification Process, *Bull. Geol. Soc. Am.*, 1993, vol. 105, no. 2, pp. 213–230.

Kaz'min, V.G., Collisions and Rifting in the Tethys Ocean History, *Geotektonika*, 1989, no. 5, pp. 14–23.

Kimura, G., and Kensaku, T., Collision, Rotation, and Back-Ark Spreading in the Region of the Ohotsk and Japan Seas, *Tectnonics*, 1986, vol. 5, no. 3, pp. 389–401.

Kimura, G., Takahashi, M., and Kono, M., Mesozoic Collision-Extrusion Tectonics in Eastern Asia, *Tectonophysics*, 1990, vol. 181, nos. 1–4, pp. 15–23.

Klitgord, R.D. and Schouten, H., Plate Kinematics of the Central Atlantic, *The Geology of North America: The Western North Atlantic Region*, Vogt, P.R. and Tucholke, B.E., Eds., Boulder : Geol. Soc. Am., 1986, pp. 351–378.

Knipper, A.L. and Sharas'kin, A.Ya., Correlation of Tectonic Events in the Mesozoic History of the Northeastern Mediterranean, *Geotektonika*, 1995, no. 1, pp. 12–22.

Livaccari, R.F., Role of Crustal Thickening and Extensional Collapse in the Tectonic Evolution of the Sevier-Laramided Orogeny, Western Unites States, *Geology*, 1991, vol. 19, no. 11, pp. 1104–1107.

Lordkipanidze, M.B., *Al'piiskii vulkanizm i geodinamika tsentral'nogo segmenta Sredizemnomorskogo skladchatogo poyasa* (Alpine Volcanism and Geodynamics of the Central Segment of the Mediterranean Foldbelt), Tbilisi: Metsniereba, 1980.

Maruyama. Sh. and Seno, T., Orogeny and Relative Motions: Example of the Japanese Islands, *Tectonophysics*, 1986, vol. 127, nos. 3/4, pp. 305–329.

Mitrofanov, N.P., The Vatyn Nappe of the Central Koryak Folded Zone, *Geol. Geofiz.*, 1977, no. 4, pp. 28–35.

Murdmaa, I., Nemliher, J., Bogdanova, O., *et al.*, Ferromanganese and Phosphatic Hardgrounds on the Western Pacific Guyots Drilled during Legs 143 and 144 ODP, *Proc. ODP Sci. Results*, College Station: Ocean Drill. Program, 1995, vol. 144, pp. 419–428.

Okada, H. and Sakai, T., Nature and Development of Late Mesozoic and Early Cenozoic Sedimentary Basins in Southwest Japan, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1993, vol. 105, no. 5, pp. 3–16.

Plafker, G., Nokleberg, W.J., and Lull, J.S., Bedrock Geology and Tectonic Evolution of the Wrangellia, Peninsular, and Chugach Terranes along the Trans-Alaska Crustal Transect in the Chugach Mountains and Southern Copper River Basin, Alaska, *J. Geophys. Res.*, 1989, vol. 94, no. 4, pp. 4255–4295.

Pushcharovskii, Yu.M., Tectonics of the Indian Ocean, *Geotektonika*, 1995, no. 4, pp. 73–91.

Pushcharovskii, Yu.M. and Melankholina, E.N., *Tektonicheskoe razvitie Zemli: Tikhii okean i ego obramlenie* (Tectonic Evolution of the Earth: The Pacific Ocean and Framing Structures), Moscow: Nauka, 1992.

Pushcharovskii, Yu.M., Zinkevich, V.P., and Mazarovich, A.O., Imbricated Nappe Structures of the Northwestern Pacific Margins, *Geotektonika*, 1983, no. 6, pp. 30–45.

Robertson, A. and Degnan, P., The Dras Arc Complex: Lithofacies and Reconstruction of a Late Cretaceous Oceanic Volcanic Arc in the Indus Suture Zone, Ladakh Himalaya, *Sediment. Geol.*, 1994, no. 92, pp. 117–145.

Rozhdestvenskii, V.S., Tectonic Evolution of Sakhalin, *Tikhookean. Geol.*, 1987, no. 3, pp. 42–50.

Ruzhentsev, S.V., Byalobzheskii, S.G., Grigor'ev, V.N., *et al.*, Tectonics of the Koryak Upland, in *Ocherki tektoniki Koryakskogo nagor'ya* (Tectonic Reviews on the Koryak Upland), Moscow: Nauka, 1982, pp. 136–189.

Sager, W.W. and Scotese, C.R., *Mesozoic and Cenozoic Plate Reconstructions*, New York: Elsevier, 1989.

Scotese, C.R., Jurassic and Cretaceous Plate Tectonic Reconstructions, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 1991, vol. 87, pp. 493–501.

Sengor, A.M.C. and Yilmaz, Yu., Tethyan Evolution of Turkey: A Plate Tectonic Approach, *Tectonophysics*, 1981, vol. 75, pp. 181–241.

Taira, A., Pre-Neogene Accretion Tectonics in Japan: A Synthesis, *Lithosphere Studies in and around Japan*, Akomoto, S., Uyeda, S., and Kono, M., Eds., Tokyo: DELP, 1985, pp. 51–63.

Zharkov, M.A., Murdmaa, I.O., and Filatova, N.I., The Initial Early Cretaceous Paleogeography, *Stratigr. Geol. Korrelyatsiya*, 1998, no. 1, pp. 47–70.

Zharkov, M.A., Murdmaa, I.O., and Filatova, N.I., Paleogeography of the Mid-Cretaceous Period, *Stratigr. Geol. Korrelyatsiya*, 1995, no. 3, pp. 15–41.

Zinkevich, V.P., *Formatsii i etapy tektonicheskogo razvitiya severa Koryakskogo nagor'ya* (Formations and Stages in Tectonic Development of the Northern Koryak Upland), Moscow: Nauka, 1981.