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Cretaceous Environments of Asia

Edited by
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Cretaceous Environments of Asia

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DEVELOPMENTS IN PALAEONTOLOGY AND STRATIGRAPHY, 17

Cretaceous Environments of Asia

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Preface

As the International Geological Correlation Programme project 350 "Environmental and Biological Change in East and South Asia during the Cretaceous", draws to a close, we present a synthesis of some of the principal environmental characteristics of the Cretaceous System in East and South Asia. This synthesis was outlined at the Fifth Meeting of the IGCP 350 Regional Coordinators in Jabalpur and Ahmedabad, India, December, 1997, and was strongly endorsed by all regional coordinators.

The initial research plan of IGCP 350 was to characterise the biological, climatological, and physical Cretaceous environments of Asia, integrated by paleogeographical regions in order to better understand the entire development of the region during this pivotal time. However, given the development of this project, the research is summarised by country, although we have tried to emphasize the natural Cretaceous regions and the changes in the biological, climatological, and physical environments within them. This has involved strati- graphy, sedimentology, paleontology, geochemistry, tectonics, petrology, mineralogy, and geophysics to confirm the synchroneity of environmental change events (see Okada, this volume). The papers in this volume cover Far East Russia, Mongolia, eastern China, Korea, Japan, Philippines, Viet Nam-Laos-Cambodia, Thailand, and India. While these countries do not encompass the entire region anticipated by this project, we are confident that this coverage is sufficient to provide a good perspective of the evolution of the region during the Cretaceous in addition to presenting new data. The Cretaceous record provides a wealth of marine and nonmarine data on its climate, biotic diversity, ocean chemistry, ocean circulation, and ocean fluctuation, as well as the fundamental plume tectonism that appears to have driven much of the environmental change in this broad region, including an enhanced greenhouse effect and high sea levels.

This project was born out of a conviction that a relatively complete scenario of Cretaceous environmental change could be identified in the continental continuum from India to the Russian Far East and in adjacent regions such as Australasia. This area also provides a rich source of information to record the interaction between the Tethys, the pan-Pacific realm, the Boreal, and northern Pacific regions.

The papers assembled here have achieved a great deal with respect to the environmental issues as defined above, and do so in English in an effort to bring this research to a broader international audience than has been the case hitherto. Kirillova et al. summarise the Cretaceous environments in Far East Russia with much new data from a vast territory. Khand et al. present a useful review of the Cretaceous System of Mongolia based on detailed analyses of fossil assemblages and the impacts upon them by the changing environment. Chen identifies paleogeographic and environmental changes during the Cretaceous in eastern China, and provides comments on remarkable dinosaur fauna from northeastern China. Chang et al. link the stratigraphic record of the Upper Mesozoic in the Korean Peninsula with active magmatism.

Okada and Sakai summarise new Cretaceous data that relate basin formation and tectonics in Japan. Hirano et al. analyse ammonoid diversity and show its relation to marine anoxic events in Japan. Kimura summarises climatic provinces of eastern Asia based on fossil land plants in Japan and adjacent regions. Kinoshita describes the Southern Pacific Superplume in relation to the Cretaceous active margin in the eastern Eurasia continent. Maekawa analyses the genesis and tectonic setting of Jurassic-Cretaceous high pressure metamorphic belts in circum-Pacific regions.

Militante-Matias et al. summarise the Philippine Cretaceous and its environments, and Vu Khuc presents recent research on the Cretaceous System of Viet Nam, Laos and Cambodia. Meesook gives a concise review of the Cretaceous sequence in the Khorat Plateau, northeastern Thailand with well-organized accounts on the paleoenvironmental analysis. Tandon synthesizes spatio-temporal patterns of environmental changes in Cretaceous aulacogen sediments in the Gondwana region of Central India.

Finally, Okada presents an overview of the scientific achievements in IGCP 350 regarding 27 topics related to enviromental problems in the Asian region as manifested by stratigraphic correlation, plume-related events, biological and climatic events, and the regions natural resources.

We would like to express our sincere appreciation to UNESCO/IGCP, the National Working Groups of the participating countries, and to the Regional Coordinators for their critical support and enthusiams. Okada wishes to thank the following organizations and individuals for their help and support in many ways: Ministry of Education, Science, Culture and Sports (Monbusho) of Japan, IGCP National Committee of Japan, Japan Petroleum Exploration (Japex) Co., Ltd., and Oyo Corporation; Drs. T. Sakai and F. Nanayama, Ms. E. Goya and S. Hayakawa on the Secretariat, Professors Emeriti T. Matsumoto, Fellow of the Japan Academy and Y. Takayanagi. Mr. T. Kaneda of the Oyo Corporation Kyushu Branch, provided significant technical help and editorial advice. This publication has been made possible by contributions by many colleagues throughout eastern and southern Asia, to whom we are very grateful.

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Cretaceous Environments of Asia
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Cretaceous environmental changes of East Russia

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The present paper summarizes the results of the investigations carried out in East Russia within the framework of the International Program of Geological Correlation (350) supported by UNESCO entitled "Cretaceous Environmental Change in East and South Asia" (1993-1998). Multidisciplinary studies made by paleontologists, biostratigraphers, sedimentologists, paleovolcanologists, and tectonists permitted the determination of synchroneity and interrelation of different geological and biological events, and re-establishment of the succession of Cretaceous climatic, biological, and physical-geographical changes in East Russia.

In the conditions of an active continental margin permanently existing during the Cretaceous, and being displaced eastward, three major events, which affected essentially environmental and biotic changes, have been distinguished. The principal events occurred in the Hauterivian, when due to the plate reorganization and oblique subduction, a regime of transform margin was dominant, a transregional regression occurred, and the first Angiospermae appeared.

The second tectonic reorganization occurred in the mid-Albian resulting in the formation of a giant East-Asian volcanic belt, which formed a longitudinal zonality against the background of the latitudinal zonality. A total cooling and considerable biota rejuvenation were caused by this event.

The third considerable reorganization occurred in the mid-Maastrichtian that was characterized by the elevation of the continental margin, fall in temperature, and a cardinal reorganization of ecosystems with the extinction of numerous flora and fauna, including dinosaurs.

We thank N.J. Mateer for critical review of the manuscript. The work of the Russian Working Group was supported by National Committee of the Russian Geologists, Heads of the Institutes, and the Presidium of the Far East Branch of the Russian Academy of Sciences, and Prof. H. Okada, the Leader of the Project IGCP 350.

1. INTRODUCTION

Cretaceous deposits of East Russia occupy a broad area extending from Chukotka to Primorye, a distance of more than 4500 km. The Cretaceous environment was diversified and inherited the main features of that of the Upper Jurassic. Principal structural constituents in the present structural pattern are the Siberian (or East Asia) craton defined from the east by a miogeosynclinal fold belt, and a system of smaller ancient blocks (the Stanovoy, Bureya, Khankai blocks), and a collage of terranes that attached to the craton from the east and southeast at different times [1, 2, 3, 4]. This diverse structural environment caused variability of Cretaceous landscapes and related sedimentary systems (Figure 1). Coastal lowlands with well-developed river networks, oxbow-lakes, lakes and swamps adjoined the paralic basins, which existed permanently during the Cretaceous and migrated mostly eastward. During certain periods in the Cretaceous these lowlands were covered with a shallow sea that abounded with islands. In the hinterland, there were located uplifted areas and plains, where continental coal-bearing deposits accumulated in isolated basins. During the Cretaceous terrigenous sedimentation, together with volcanogenic materials, predominated. Siliceous and siliceous-clayey deposits typical of a back-arc basin and ocean are restricted within their distribution.

Through the Cretaceous, latitudinal climatic zonality prevailed while longitudinal zonality was manifest along the continent-ocean boundary.

Fundamental features of the Cretaceous deposits of East Russia have been studied by many workers, with considerable advances made in the last decade, thanks to the work by about 60 geologists on the projects of the International Program of Geological Correlation (245, 350) and Global Sedimentation Geological Program [5]. A complete list of publications in 1960-1993 has been compiled [6], but further references will be principally to more recent publications and materials obtained in the course of the work on IGCP Project 350.

2. MAJOR CRETACEOUS STAGES AND EVENTS

2.1. Physical-geographical and geodynamical environment

2.1.1. Early Cretaceous

Cretaceous deposits of East Russia have traditionally been divided into four regions [7]: East Siberia, Northeast, Far East and the Pacific coastal region (Figure 1).

Lower Cretaceous deposits of East Russia are divided rather distinctly into pre-Hauterivian and post-Hauterivian parts. During the Hauterivian, a Kolymian folding phase occurred throughout most of East Russia and left-lateral strike-slip movements became more active.

The Lower Cretaceous history of East Russia closely follows that of the Upper Jurassic. The coastline of the East Siberian region (the North Asian or Siberian craton) migrated to 300-500 km north and east during the Berriasian and a huge coastal plain was thus created and was bordered with low hills from the east. Under these continental conditions there accumulated terrigenous coal-bearing formations about 1500 m thick, one of the greatest in terms of all geological epochs on the Earth. Volcanogenic sedimentary formations of about 600 m thick developed in the southern part of the region (Figure 1).

The Northeastern region is characterized by a complex evolutionary history that determined the diversity of geological environments. According to the sedimentary peculiarities and character of the stratigraphic sequence, three types of sections are distinguished. In the

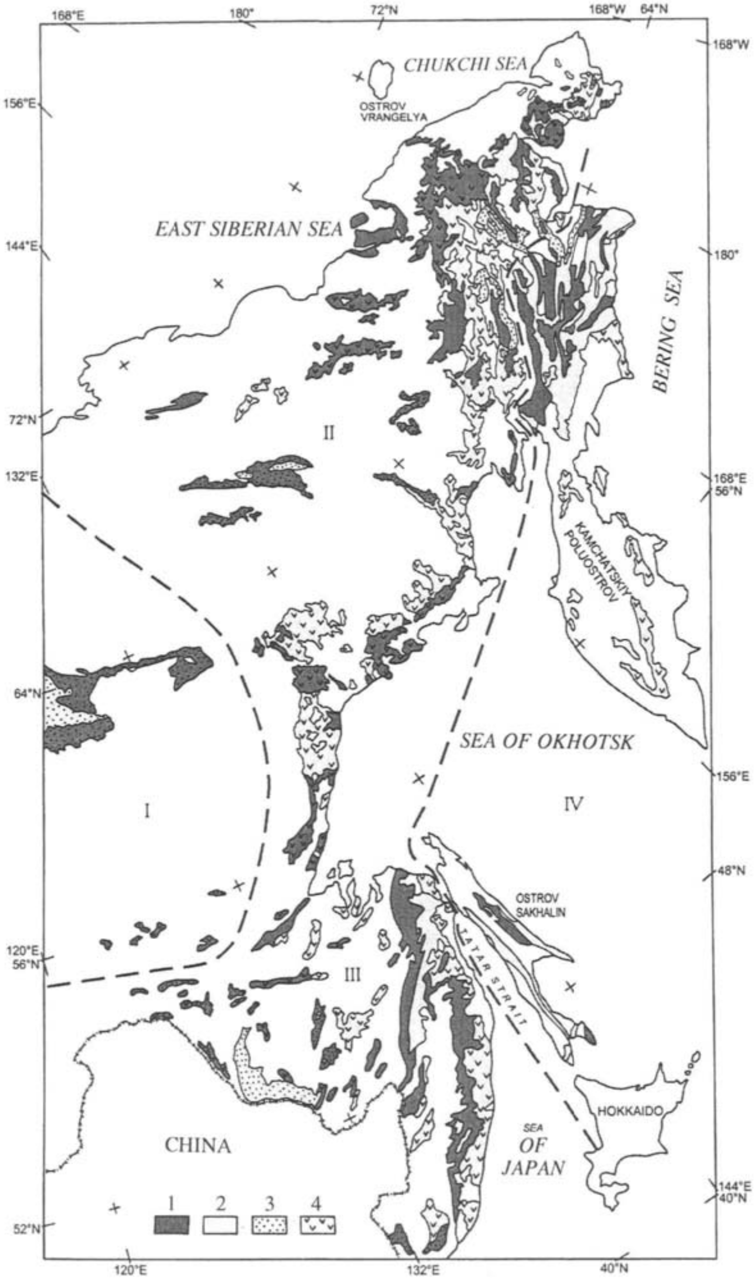


Figure 1. Distribution of Cretaceous sediments in East Russia. 1, 2 - marine: 1 - Lower Cretaceous; 2 - Upper Cretaceous. 3, 4 - continental: 3 - terrigenouscoal-bearing; 4 - volcanogenic. Regions: I - East Siberian; II - North Eastern; III - Far Eastern; IV - The Pacific coast.

southwestern part of the region two large emergent lowland areas, the Zyryanskaya and Priokhotsky, separated by low hills, in which continental coal-bearing terrigenous formations (about 1500 m thick) accumulated. In Priokhotie, Berriasian-Hauterivian intermediate-basic volcanics, their tuffs, volcanogenic-sedimentary rocks were deposited. The northeastern part of the region appears to have been a marine basin with archipelagoes, which deepened southeastward. The most investigated section of the Berriasian-Valanginian marine terrigenous deposits, about 1000 m thick, is reported from the Oloyisky zone. It is considered a Late Jurassic-Early Cretaceous Oloyisky-Svyatonossky island arc with an adjacent fore-arc basin. To the northeast, terrigenous sedimentation to a thickness of 2000 m occurred in the South Anyui basin. Regions with a lesser sediment thickness and numerous depositional hiatuses form a third type of the section.

One of the geodynamic interpretations of pre-Hauterivian period in the development of the Northeastern region is described by Parfenov [8]. Due to the opening of the South Anyui basin (Figure 2a) in the Late Jurassic-Neocomian, a collision occurred between the Kolyma-Omolon superterrane and the North Asia craton [8]. Compositionally and genetically different terranes of the Kolyma-Omolon superterrane were amalgamated into one tectonic unit at the mid-upper Jurassic boundary prior to its accretion to the North Asian craton, which occurred in the Late Jurrasic. All these terranes are overlain unconformably by Upper Jurassic to Cretaceous volcano-sedimentary rocks. Along the northern margin of the Kolyma-Omolon superterrane, the Oloy-Svyatonossky volcanic arc, composed of calc-alkaline volcanics, is distinguished. Fore-arc turbidities are known from the South Anyui terrane. South of the Kolyma-Omolon superterrane, the Koni-Murgalsky Triassic-Neocomian island-arc terrane is distinguished. Post-accretion formations are represented by Jurassic-Neocomian volcano-sedimentary depo- sits of the Udsky belt. It is a marginal-continental Andian type magmatic arc, which is the southwestern continuity of the Udsk-Murgal island arc.

During the Hauterivian-Albian the Lena and South Yakutiya coal fields continued to form in the East Siberian region, where about 1500 m terrigenous coal-bearing formations were deposited. In the southern part of the East Siberian region within a restricted area volcanogenic sedimentary deposits, about 500 m thick, continued to accumulate with hiatuses.

Two principal zones of sedimentation remained in the Northeastern region despite the Hauterivian regression. In the central part of the region, the Momo-Zyryansky and Sugoisky basins developed through the mid-Albian and were bordered by mountains and filled with terrigenous coal-bearing deposits of about 7000 m thick. To the east of the early Neocomian seas, only in the Anyui sea gulf sediments accumulated through the Hauterivian-Aptian (Figure 3): sandstones, siltstones, conglomerates, about 2000 m thick, with lenses of coal and tuffs of andesites. Albian deposits are represented by continental facies. The Aptian-Albian boundary is identified by the change from marine to continental deposits. The lower Albian is composed of sandstones, conglomerates, siltstones of about 1200 m thick, while the upper part is dominated by unconformably laid volcanics of about 900 m thick.

According to the geodynamic reconstructions made [8], the closure of the South Anyui basin was caused by the opening of the Canadian basin (Figure 2b, c), and at which time the Chukchi terrane, previously situated along the northern margin of North America, became attached to Siberia in pre-Albian time. The Koni-Murgal arc was also accreted at that time. As suggested earlier, terranes migrated from the southeast, from the Pacific, but based on the latest paleomagnetic data derived from different parts of the Koni-Murgal arc, it has been proposed to explain the mid-Jurassic-Cretaceous tectonics of Northeastern Russia in terms of

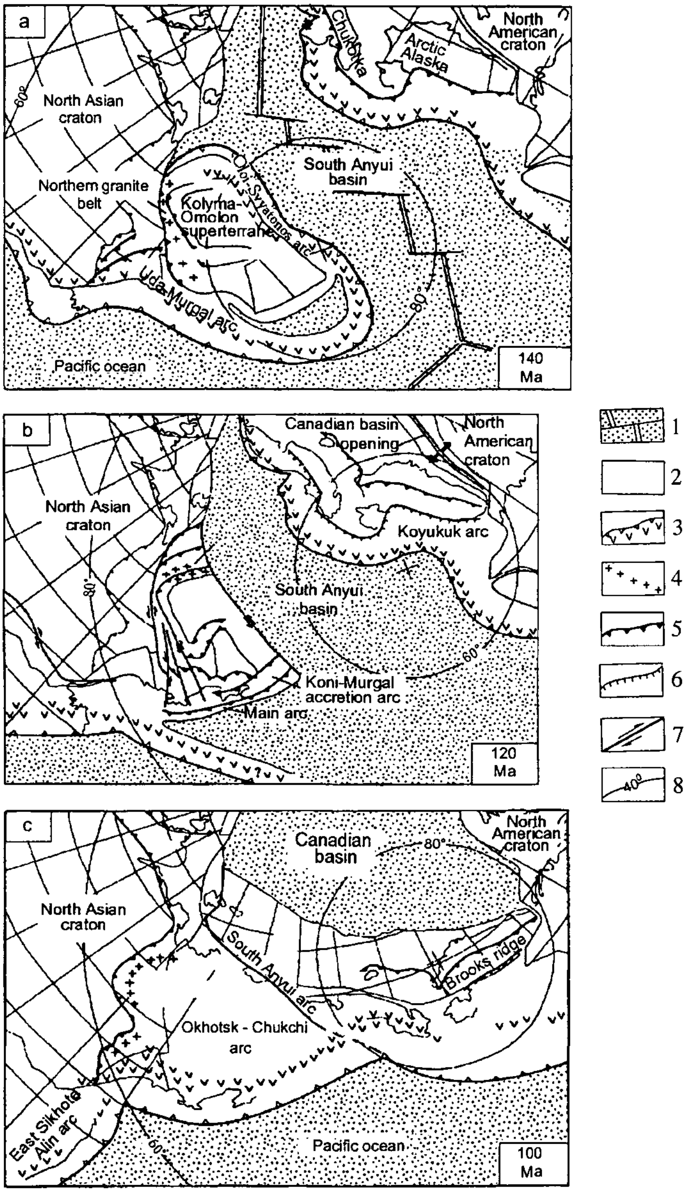


Figure 2. Three Late Mesozoic paleotectonic reconstructions (from Parfenov [8]): a - 140 Ma (Late Jurassic/Cretaceous boundary); b - 120 Ma (Hauterivian); c - 100 Ma (Albian).
1 - basins with oceanic crust; 2 - continents and microcontinents (terranes with continental crust) with the modern coordinate set; 3 - magmatic arcs; 4 - collision granite belts; 5 - overthrusts; 6 - faults; 7 - strike-slip displacements; 8 - paleolatitudes.

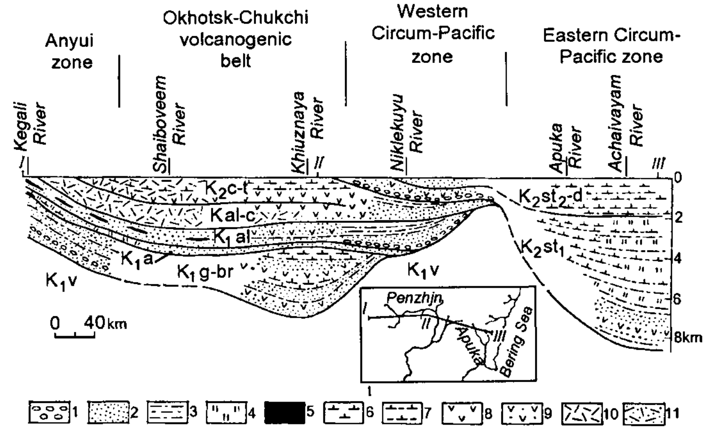


Figure 3. Lithological section of the Hauterivian-Danian sediments from the Omolon river basin to the Achaivayam river basin, North Eastern region (from Krasny and Putintsev [7]). Legend: 1 - conglomerates; 2 - sandstones; 3 - siltstones and mudstones; 4 - cherts; 5 - coals; 6 - basic volcanics; 7 - basic tuffs; 8 - intermediate volcanics; 9 - intermediate tuffs; 10 - acid volcanics; 11 - acid tuffs.

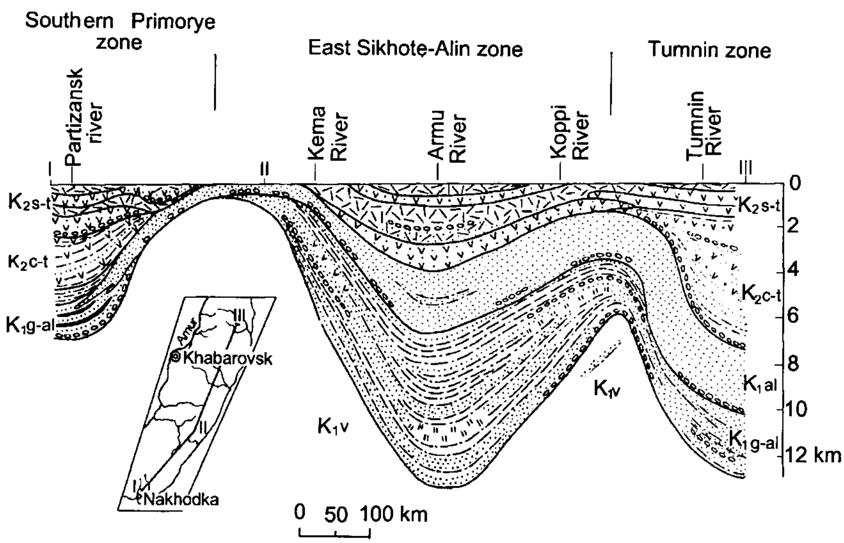


Figure 4. Lithological section of the Hauterivian-Senonian sediments of the East Sikhote-Alin, Far Eastern region (from Krasny and Putintsev [7]). See the legend in Figure 3.

the Eurasian and North America plates interaction, when the Atlantic ocean was opening [9]. Terranes are considered to pass through the pole during the Jurassic and in the post mid-Jurassic they moved southwards.

The development of the Far Eastern region during the Early Cretaceous closely follows that of the Late Jurassic. There existed three types of basins in the Early Cretaceous. The Lower Amur and Sikhote-Alin basins located eastward are regarded as the first type, and appear to be rather deep seas, in which in the conditions of stable downwarping environment mostly Berriasian to Valanginian turbidites, about 6000 m thick, accumulated [7, 10, 11] with a subordinate amount of cherty-volcanic rocks and limestones. Horizons with olistostromes containing various Paleozoic to Mesozoic clastics are often observed near the continent boundary [2,12,13]. Some investigators suggest the existence of an Early Cretaceous oceanic basin, the deposits of which (jasper, cherts, mudstones) are associated with alkaline basalts described in the Kiselyovka-Manominsky terrane [14].

Basins located along the margin of the pre-Jurassic continental structures (Toromsky, Udsky, Bureya, Okrainsky, South Primorsky) [7,11] comprise the second type. They represented epicontinental seas with rich benthos, in which diversified terrigenous material generally including products of the terrestrial volcanism, accumulated during the Berriasian to early Hauterivian. Here intensive downwarping epochs were repeatedly superseded by uplifting and erosion and continental sedimentation including coal formation. The thickness of the deposits varies from 2000 to 4000 m.

The third basin type comprises intracontinental rift and postcollision basins separated by low mountains (the Amur-Zeya, Pre-Dzhugdzhursky), in which continental coal-bearing volcanogenic-terrigenous deposits, about 1000 m thick, accumulated during the Berriasian-Hauterivian.

According to geodynamic reconstructions [4, 15, 16], there existed in the Early Cretaceous the Khingan-Okhotsk active continental margin bounded by the Amur suture from the east. It comprised a magmatic arc marked by a chain of volcanic areas extending from the Lesser Khingan to the Sea of Okhotsk, and an accretionary complex made up mainly of turbidites. Olistostrome horizons discovered at different Early Cretaceous stratigraphic levels are confined to zones of the largest bed-by-bed disintegrations. As a result of the oblique subduction of the Izanagi plate, the Valanginian sinistral margin-slip movements initiated in the East Asian margin [16, 17] and formation of the transform continental margin began [18]. In the eastern part of the region the Anyui microcontinent accreted [19]. These movements accompanied by local collision events became more intensive in the Hauterivian and caused uplifting of large blocks, hiatuses in sedimentation, and considerable changes in the coastline.

Marine Hauterivian deposits are known only in the southeastern Primorye, in which a continuous Berriasian-Albian section has been described [10], and no deposits are observed in the rest of the area.

During the Barremian to mid-Albian, the following four sedimentation areas can be distinguished: West Sikhote-Alin, East Sikhote-Alin, marginal-continental, and intracontinental basin areas.

The West Sikhote-Alin basin is viewed as a pull-apart sea basin filled with late Hauterivian to Barremian turbidites. In the Aptian-early Albian in the eastern part of the basin the Samarga and Udyl island arcs generated and the basin assumed the features of a back-arc basin. Among the terrigenous material volcanogenic sandstones, tuffs, and tuffaceous siltstones are predominant. The sedimentation rate sharply increased after the Aptian when island-arcs were becoming larger. The thickness of late Hauterivian-mid-Albian deposits attains 5500 m.

The East Sikhote-Alin basin (Figure 4) is interpreted as a pull-apart basin being filled with terrigenous turbidites during the Hauterivian to Barremian. Beginning in the Aptian, a mixture of the volcanic material appears in the turbidites, the rate of sedimentation increases sharply. From that time the basin can be considered as a forearc basin where sedimentation continued until the mid-Albian. The thickness of the deposits was about 9000 m.

The marginal-continental basin area covered the Udsky, Toromsky, Bureya, Razdolnensky and Partizansky basins. Terrigenous continental deposits of about 1500 m thick predominated. The deposits of the first two basins consist of mostly conglomerates and diversified sandstones (about 600 m) that contain volcanic material that is evidence for the proximity of the volcanic arcs. In the late Aptian to mid-Albian the Okhotsk-Chukchi volcanic belt and its fragments began to form along the basin margins. Rhyolites and dacites were the first volcanics to appear.

The Bureya, Razdolnensky and Partizansk basins in the Hauterivian to mid-Albian represent coastal-marine marsh plains. A cyclic terrigenous coal formation reaching a thickness of 1500 m accumulated there. The Razdolnensky and Partizansk basins are related to the pull-apart basins as their formation in the Hauterivian is associated with the intensive left-lateral strike-slip faults [20]. During the mid-Albian transgression, the sea penetrated briefly the Bureya and Partizansky basins as confirmed by the beds with marine *Trigonia* and foraminifers. From the late Albian, however, the type of sedimentation in the Partizansk and Razdolnensky basins changed sharply: after a short interval about 3500 m of continental coarse clastic, variegated volcanogenic terrigenous rocks were deposited.

The Amur-Zeya basin is related to the intracontinental basin area that expanded substantially during the Hauterivian to early Albian thus marking a post-rift subsidence stage of the development [21], during which time 1200 m of terrigenous coal-bearing deposits accumulated. Along the faults on the intracontinental uplifts, intermediate and acid volcanics erupted.

The principal structural constituents of the Far Eastern region are shown in Figure 5. In accordance with one paleotectonic reconstruction [3], Early Cretaceous sedimentation occurred in the more southerly latitudes. Only in the Albian that the Khabarovsk, Amursky and Kiselyovka-Manominsky subduction-accretion complexes of the Khingan-Okhotsk active continental margin along the sinistral faults, attached to the pre-Cretaceous continent. Others do not suggest the displacements to be so considerable [16, 22] and the events developed in different modes. Undoubtedly, however, the environment changed abruptly from the mid-Albian. After collision of a number of terranes (the Taukha, Zhuravlevka, and Kema terranes), most of the Far Eastern territory experienced uplift and intensive volcanism. Only a narrow sea inlet remained along what is today the Amur river valley.

The Pacific coast region embraces the Koryak upland, the Kamchatka peninsula and the Sakhalin island (see Figure 1). Lower Cretaceous rocks there are similar to those of the Upper Jurassic that formed during intensive subsidence resulting in accumulation of thick volcanogenic-cherty-terrigenous marine deposits, which were poor in organic remains. The following three Neocomian zones are recognized within the region: the Sakhalin, Taigonos-Anadyr and Talov-Mainitskaya [7].

In Sakhalin up to 4000 m Upper Jurassic-Lower Cretaceous deposits occur in both the western and eastern parts of the island. The lower part is composed mainly of volcanogenic-cherty rocks, in which Early Cretaceous radiolarians have been discovered [23], and occa-

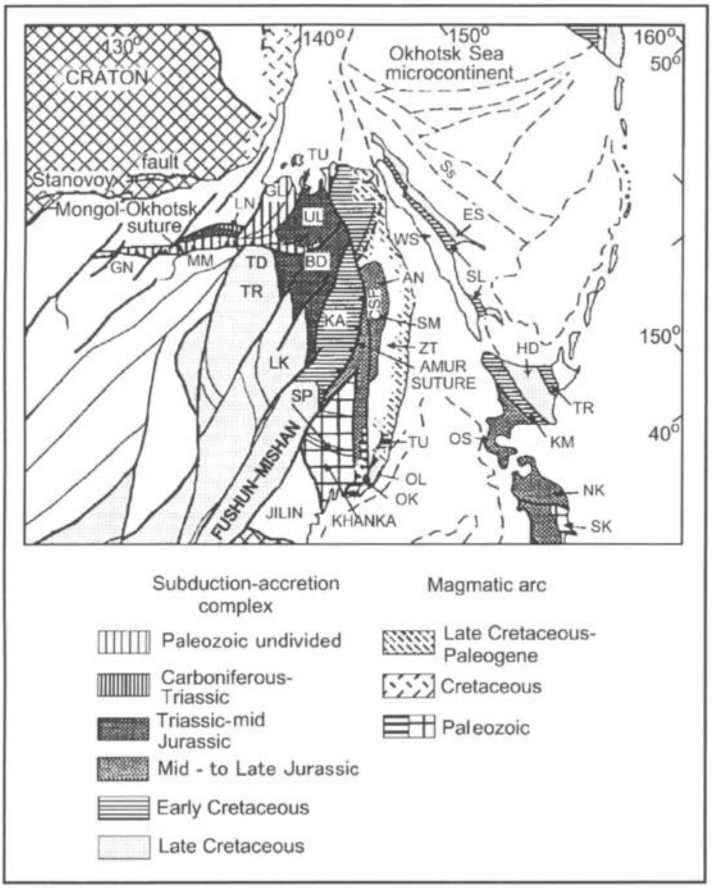


Figure 5. Generalized tectonic map of the Far Eastern region and surrounding areas (from Sengor and Natal'in [3]).
AN - Anyui microcontinent; ES - eastern Sakhalin subduction-accretion complex; GN - Gonzha block of the Tuva-Mongol massif; HD - Hidaka subduction-accretion complex; KA - Khabarovsk, Amursk, and Kiselevsk-Manominsk subduction-accretion complex of the Khingan-Okhotsk active continental margin; KM - Kamuikotan subduction-accretion complex; LK - Lesser Khingan unit; LN - Lan subduction-accretion complex; MM - Mamyn block of the Tuva-Mongol massif; NK - Northern Kitakami subduction-accretion complex; OK - Okrainsk block; OL - Olginsk subduction-accretion complex; SK - Southern Kitakami magmatic arc; SL - Susunay-Langeri subduction-accretion complex; SM - Samarka subduction-accretion complex; SP - Spassk suture; TD - Tukuringra-Dzhagdi subduction-accretion complex; TR - Tokoro subduction-accretion complex; TR - Turan unit; TU - Tau- kha subduction-accretion complex; UL - Ulban subduction-accretion complex; WS - West Sakhalin forearc basin; ZT - Zhuravlev-Tumnin foredeep basin.

sional limestone lenses with corals and foraminifers [24]. The upper part is dominated by terrigenous deposits with subordinate amounts of volcanogenic-cherty rocks. This rock unit agrees well with the Sorachi and Lower Yezo Groups on Hokkaido Island. They are approximately dated as Berriasian to Barremian [25]. Sediments of the Sorachi Group accumulated in the oceanic basin, whereas those of the whole Yezo Supergroup were concentrated in the forearc basin [26]. However, the view on the onset of the terrigenous sedimentation both in Sakhalin and Hokkaido still remains in dispute.

On the Moneron Island a Lower Cretaceous volcanogenic-sedimentary formation related to the Rebun-Kabato volcanic arc has been penetrated by boreholes, and is presumably linked to the Samarga arc located to the north [27]. The Taigonos-Anadyr zone is located between the Verkhoyansk-Kolyma and Koryak-Kamchatka fold systems providing a diverse types of sections. A typical section of the Taigonos peninsula deposits distributed in the northwestern part, has been studied on its western coast [7]. It is an upper Volginian-Valanginian series of coastal-marine and continental deposits, about 5000 m thick, composed of andesite lavas, basalts, ignimbrites, rhyolites lavas and their tuffs, which are typical volcanic arc facies. In the southeast, a typical section has been described in the southeastern Taigonos peninsula. Its lower part consists of spilites, diabases, cherty-clayey shales, turbidites, and the upper part consists of turbidites that contain horizons of olistostromes and ultramafic olistoliths. A total thickness of the section attains 6000 m.

The Koryak-Kamchatka zone is characterized by about 2500 m thick sections of Upper Jurassic (Volginian), Berriasian and Valanginian marine deposits that occur unconformably. Two section types are known, either mostly terrigenous with siltstone containing *Buchia* sp., or of radiolarian cherty-volcanogenics with interbedded limestones. In the present-day structural pattern, these sections are composed of narrow slabs, imbricated thrusts and shifts and appear to be subduction-accretionary complexes and magmatic arcs with forearc basins (Figure 6) [3].

During the Hauterivian, marine basins remained only in the regions of stable downwarping, in which mostly terrigenous marine deposits with subordinate quantity of volcanics with occasional continental intercalations accumulated. The most typical section has been described in the Penzhin river basin. Hauterivian-lower Barremian deposits comformably overlying Valanginian complexes consist of sandstones, turbidites with conglomerate beds, volcaniclastics and basalts (300-700 m thick). In the Barremian-lower Aptian deposits, an essential mix of volcanogenic component of about 1300 m thick, occurs that indicates volcanic arc activity. Predominant were tuffites and tuffs of intermediate and acid effusives containing beds of tuff breccia and tuff conglomerates. Upper Aptian deposits are represented by mostly 2200 m turbidites. Albian deposits are also dominated by turbidites with the interbeds of tuffs, conglomerates, and limestones.

Barremian to mid-Albian deposits in the western Kamchatka are represented only by about 500 m volcanogenic terrigenous complexes. In Sakhalin there is a thick sequence of deep-marine volcanogenic-cherty-terrigenous deposits that contain few organic remains. Jurassic-Early Cretaceous radiolarians are recognized at the base of the section, whereas Albian-Cenomanian radiolarians are discovered in the upper part of it .

In the Hauterivian to mid-Albian sequence of the Pacific coast region, widespread lower to mid-Albian deposits were dominated by turbidites. Upper Albian deposits in most sections occur unconformably with conglomerates at their base indicating a significant change in the tectonic regional framework.

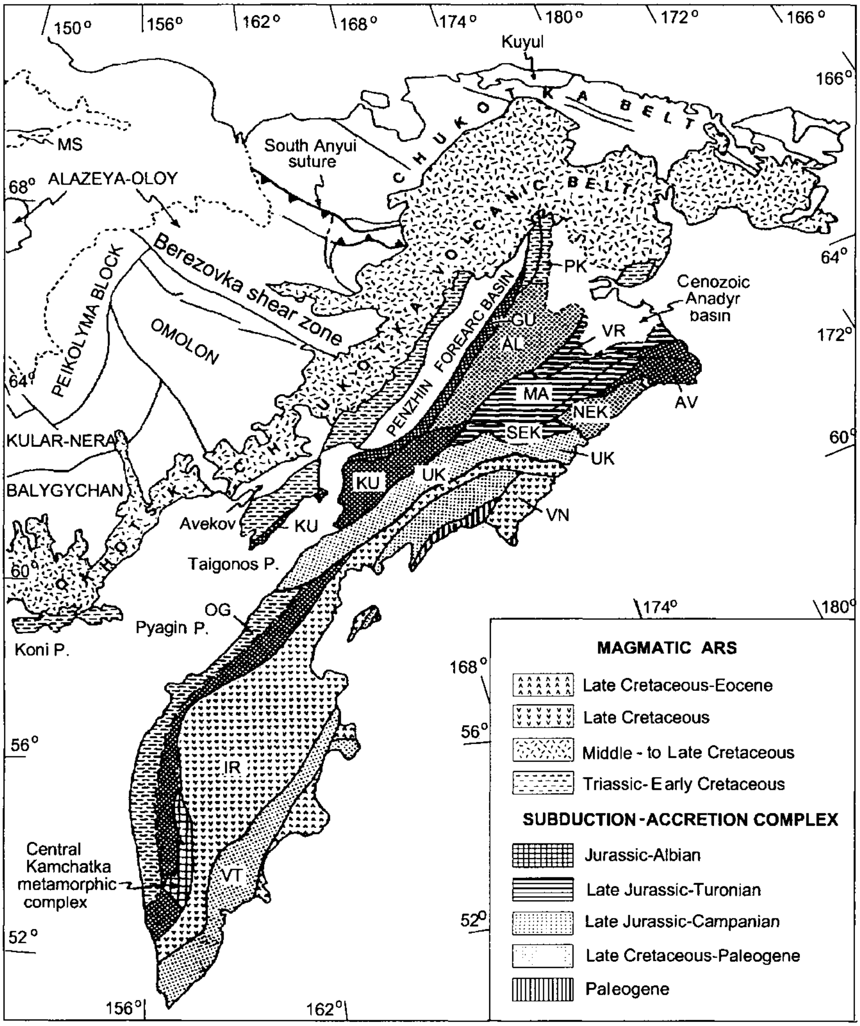


Figure 6. Generalized tectonic map of the Kamchatka-Koryak sector and surrounding areas (from Sengor and Natal'in [3]).

AL - Algan subduction-accretion complex; AV - Alkatvaam subduction-accretion complex; GU - Ganychalan-Ustbel subduction-accretion complex; IR - Iruney magmatic arc and subduction-accretion complex; KU - Kuul subduction-accretion complex; MA - Mainits subduction-accretion complex; NEK - northern Ekonay subduction-accretion complex; OG - Omgon subduction-accretion complex and Mesozoic magmatic arc; SEK - southern Ekonay subduction-accretion complex; VR - Velikorechensk subduction-accretion complex; VT - Vetlov subduction-accretion complex.

Early Cretaceous Pacific coastal region was an accretionary wedge of the Udsk-Murgal and Okhotsk-Chukchi (from mid-Albian only) magmatic arcs (Figure 2), which subsequently accreted to the East Asian craton [1, 3, 28]. By the mid-Albian, the Ganychalan-Ust-Belsky subduction-accretionary complex contiguous to the Penzhin forearc basin was formed and accreted (Figure 6), and in West Kamchatka the Omgon subduction-accretionary complex and a magmatic arc were formed. Oblique subduction was accompanied by dextral displacement and compression during the Late Jurassic-Early Cretaceous [29]. The Chukchi-Alaska block accretion to the East Asian craton was followed by gabbro-plagiogranite and diorite-granodiorite emplacement.

The formation of the extended Okhotsk-Chukchi magmatic arc (or volcanic belt) initiated in the mid-Albian [30, 31] and was the most remarkable event at the Early/Late Cretaceous transition. The East Sikhote-Alin belt, which also began to form in the mid-Albian, represented the southern continuation of the Okhotsk-Chukchi belt.

2.1.2. Late Cretaceous

After the mid-Albian, seas retreated eastward, outside the volcanic arcs. Along the Siberian craton margin in the East Siberian region Late Cretaceous cross-bedded sandstones, about 800 m thick, and pebbles interbedded with lignite, continued to accumulate. To the south, volcanics were accumulating. Acid volcanics (about 200 m thick) dominated during the early Cenomanian, whereas in the late Cenomanian-Turonian intermediate and basic volcanics of the same thickness occurred.

In the Northeastern region the extensive Okhotsk-Chukchi arc (or belt) continued to form in the Late Cretaceous. Its emergence was associated with the subduction of the Kula plate under the continental margin of Asia, which superseded the mid-Albian collisional regime. The belt overlapped the thrust-fold structures of the Verkhoyansk-Chukchi tectonic area, Koryak-Kamchatka orogenic belt and collisional sedimentary basins. Within the belt a 200-5000 m thick basalt-andesite-dacite-rhyolite association of calc-alkaline series was deposited [30, 31]. The belt evolution is distinguished by early and late stages, divided at the Albian-Cenomanian boundary. About 90 % of total volume of volcanics (over 10^6 km^3) originated in the early stage [30]. In the Santonian-Campanian, volcanism largely ceased and acid volcanics occurred unconformably. Belyi [30] does not consider them to belong to the belt, but Filatova [31] suggests the Okhotsk-Chukchi volcanic belt continued to be active till the end of the Senonian. During the Maastrichtian-Paleogene, alkaline olivine plateau-basalts erupted from central type basalts, regarded by Filatova [31] as rift intra-plate basalts.

In valleys of the Okhotsk-Chukchi mountain belt continental volcanogenic-terrigenous sediments accumulated, often with coal-bearing deposits. Further west, continental sedimentation and synchronous volcanism continued only in the following basins: the Ukveemsky basin, from which 600 m of Cenomanian ignimbrites and dacite tuffs are known, the Momo-Zyryansky basin with 600 m of Turonian terrigenous rocks accumulated after a hiatus, and the Sugoisky basin, in which 3500 m of Cenomanian volcanogenic sediments occur, of which andesites, rhyolites and their tuffs were predominant.

During the Late Cretaceous the Far Eastern region contained distinct eastern and western depositional areas. The western continental area was characterized by the eruption of acid to intermediate volcanics during the Cenomanian-Turonian, but in the Amur-Zeya basin a hiatus is observed. During the Turonian-Senonian, extensive 300 m lacustrine sandstones and clays with a rich fauna were deposited. Along the basin margins conglomerates occurred with the deposits reaching a thickness of 800 m. Cyclical Maastrichtian-Danian sandstones, gravels, claystones (about 450 m thick) are widespread, with dinosaur remains occurring in mid-Maastrichtian beds.

In the eastern area, the formation of Late Cretaceous East Sikhote-Alin volcanic belt occurred in the coastal part of the continent. It consists of alternating series of lavas and intermediate to acid tuffs of about 1500 m in total thickness. During the Late Cretaceous the West Sikhote-Alin belt consisting of 1800 m of alternating series of lavas and intermediate to acid tuffs continued to form. Volcanic arcs represented topographic highs that probably had an impact on local climatic conditions and thus sedimentary regimes. In the back arcs Cenomanian-Turonian about 3000 m of continental variegated volcanogenic-sedimentary deposits were generated suggesting an initial high rate of volcanic arc denudation while forearc troughs were being formed.

During the Late Cretaceous most of the Far Eastern region was land, and only in the north-eastern part, along the present-day Amur river valley, did a marine embayment persist. It represented a back-arc basin, in which about 4000 m of coarse sandstones, siltstones, mudstones with a mix of tuffaceous material, with basal conglomerates were deposited during the Cenomanian-Turonian. At the end of the Turonian, the sea retreated from the region completely, and since that time until the latest Cretaceous, up to 1000 m of andesite and rhyolite volcanic rocks formed. The sea retreated to the eastern margin of the East Sikhote-Alin belt.

The Pacific coastal region of Russia is also divided into western and eastern areas in the Late Cretaceous. Coastal-marine, shelf, lagoon sandy-argillaceous deposits alternating with parallic coal-bearing strata of the marginal continental basins (Figure 7) are distributed in the western area. The deposits are rich in both fossil fauna and flora that provide more precise dating and correlations with nonmarine beds. The Aptian-Late Cretaceous sedimentation history in this area is distinguished by flysch and molasse stages. The former is of Aptian-Cenomanian age, and the latter is of Turonian-Maastrichtian. The second stage is subdivided into three sedimentary cycles with transgressions in the early Turonian, late Santonian-early Campanian, and the late Campanian-early Maastrichtian. Regression and nonmarine sedimentation took place in the Coniacian-early Santonian, the mid-Campanian and the late Maastrichtian [32, 33]. Correlation of the Upper Cretaceous deposits is based on the zonal stratigraphic schemes identified from ammonites, inoceramids, and other bivalves, foraminifers, and radiolarians [24].

This stratigraphy permitted correlation of Late Cretaceous coal beds in the western zone [33]. Two major coal accumulation epochs, the Arkovskian (late Turonian-Santonian), during which the thickest and numerous coal seams originated, and the Jonkierian (mid-Campanian), during which coal-bearing deposits were widespread from Japan to Chukotka.

In Sakhalin only the northern part belongs to the western area. In the generalized section of the western area the upper Albian-Turonian deposits (700-5000 m thick) contain conglomerates at the base and rest unconformably on the subjacent beds and consist of sandstones, siltstones, turbidites, tuff and conglomerate rocks. The Coniacian to lower Campanian 1000-3700 m section consists of sandstones with conglomerate interbeds. The upper Campanian sandstones and conglomerates (400-1400 m thick) frequently contain tuff and coal interbeds. Maastrichtian deposits (300-1500 m thick) are composed of sandstones with interbeds of conglomerates, siltstones, tuffs, but coals are rare. Upper Maastrichtian to Danian deposits are represented by either sandy-conglomerate coal-bearing complex of 800-1300 m thick or sand-

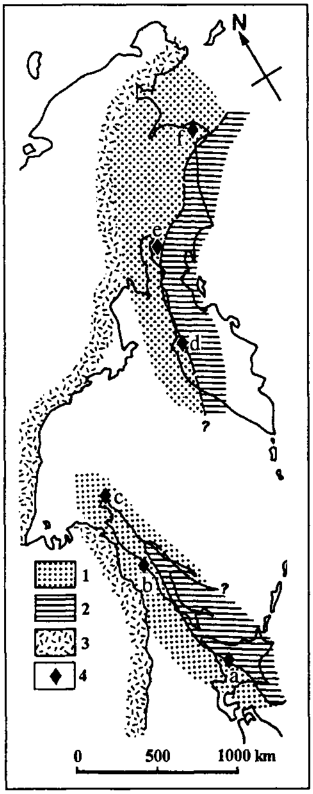


Figure 7. Location of the sections with frequent alternation of Late Cretaceous marine and nonmarine deposits (from Salnikov [33]).
1 - continental coal-bearing deposits; 2 - marine deposits; 3 - volcanic belt; 4 - main sections: a - meridional zone of Hokkaido; b - Alexandrovsk area of Sakhalin; c - Schmidt Peninsula; d - Tigil area of West Kamchatka; e - Penzhin Bay; f - Ugolny Bay.

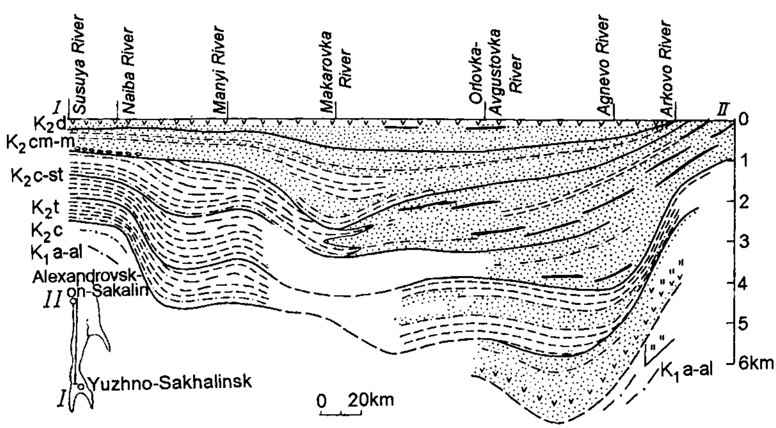


Figure 8. Lithological section of the Cretaceous deposits of the West Sakhalin Mountains [7]. See the legend in Figure 3.

stones and siltstones containing a fossil marine fauna, but no ammonite indices. These beds are practically inseparable from those of the Paleocene deposits.

In the eastern sedimentary area of the Pacific coastal region, Late Cretaceous marine deposits are widespread, but in general contain few organic remains except for southwestern Sakhalin, where the most complete Upper Cretaceous shallow-water section along the Naiba river has been described for East Russia (Figure 8). This section is well confirmed by ammonites, inoceramids, foraminifers and other organic remains [24]. In East Sakhalin a 6000 m thick Turonian-Maastrichtian series of deep-water volcanogenic-cherty-terrigenous deposits has been recorded, the upper part of which contains turbidites with olistostrome horizons. Besides, in southeastern Sakhalin island-arc trachyandesites are found. It is there that Campanian-Danian complex of coastal-marine and continental terrigenous deposits with coal interbeds is developed. The similar change in deposits can be observed on the Hokkaido Island (Campanian to Maastrichtian Hakobuchi Group).

Small outcrops occur on the islands of the Lesser Kuril Range consisting of Campanian conglomerate-breccia, lava-breccia, tuffstones with inoceramids and Maastrichtian turbidites with basalt sills, and an unconformable late Maastrichtian-Danian conglomerates, and siltstones with fauna and flora [34, 35].

Pre-Campanian volcanogenic-cherty-terrigenous deposits, the age of which has not been confirmed by fossils, are distinguished in East Kamchatka. Campanian complexes of about 400-3000 m thick superpose these deposits. Besides, they contain limestone interbeds and lenses with inoceramids. Presumably Maastrichtian deposits with a thickness of about 3000 m occur disconformably on the Campanian strata. Their lower part consists of volcaniclastics, whereas the upper part consists of cherts, sandstones and siltstones.

In southeastern Koryakia, similar Campanian to Danian cherty-volcanogenic-terrigenous deposits, 4000-12000 m thick, overlie unconformably Valanginian-Hauterivian complexes of the western Pacific coastal region. Scarcity of fossil indices and complex tectonism make the thickness estimation very approximate.

For the last 5-7 years, data on detailed stratigraphic, structural and paleomagnetic studies have been obtained and analyzed for the Cretaceous of NE Russia that considerably changed the previous concepts [1, 28, 36, 37, 38, 39, 40, 41, 42].

Tectonic and paleogeodynamic reconstructions are evidence that mid-Albian deformation created a new structural pattern characterized by its lateral environmental range and corresponding complexes (from the west to the east): volcanogenic-sedimentary complex of the Okhotsk-Chukchi volcanic belt, terrigenous shelf deposits, continental slope and bottom, volcanic-tuffaceous and hyaloclastic-basaltic island-arc series, jasper-basalt and volcanic-tuffaceous-cherty oceanic formations [28].

The expansion of the Late Cretaceous accretionary wedge in front of the Okhotsk-Chukchi arc terminated during the Late Cretaceous in the response of the collision of the continental margin with the Okhotsk Sea microcontinent located in the central part of the Okhotsk Sea [3]. Part of the microcontinent can be observed in the southwestern Kamchatka.

Eastwards, fragments of another Late Cretaceous subduction-accretionary complex and Turonian-Maastrichtian magmatic arc formed and are considered a base of the Kamchatka Cenozoic volcanic arcs [43]. The subduction-accretionary complex consisted of ophiolites, pelagic sediments and turbidites that represented a packet of slabs with southeastern vergence. Shapiro [38] and Kovalenko[39] established that during the Late Cretaceous the Achaivaam-Valaginsky Campanian-Maastrichtian intra-oceanic arc was located at 50° N, attaching to the continent only in the early Paleocene. As far as the Okhotsk-Chukchi belt concerned, the paleomagnetic data showed that it preserved the same location during the Late Cretaceous.

Another NW-SE Late Cretaceous island-arc system persisted west of the Okhotsk Sea microcontinent. It is represented by a late Albian to Paleogene East Sikhote-Alin volcanic belt, forearc trough and subduction-accretionary complex that crops out in Sakhalin. The Shmidt suture containing ophiolites and shales [32] appears to be a geo-suture, along which the Okhotsk Sea microcontinent accreted to Sakhalin during the early Paleocene.

Late Cretaceous island-arc system of NE extension existed along the southern margin of the Okhotsk Sea microcontinent [32, 44]. Fragments of the volcanic arc and forearc trough of the accretionary prism are known in southwestern Sakhalin, southeastern Hokkaido Island (Nemuro Group) and on the islands of the Lesser Kuril Range. According to geodynamic reconstructions, this system was located at approximately 50° N during the Late Cretaceous and achieved its current position along meridional dextral displacements only during the Late Cenozoic [25, 44].

2.2. Biodiversity

2.2.1. Early Cretaceous

2.2.1.1. Paleofauna

The presence of all Cretaceous stages in East Russia has been substantiated primarily by ammonites, inoceramids, buchias, aucellines, and to a lesser extent by belemnites, radiolarians, foraminifers and floral remains. A detailed biostratigraphic analysis of the Jurassic-Cretaceous boundary strata showed a discrepancy between the Boreal and Tethyan scales and the affinity of the upper Volginian substage to the Cretaceous System [45]. Nevertheless, fossil remains from all Early Cretaceous stages with the predominance of endemic species of the Pacific paleobiogeographic realm allow local biostratigraphic units to be correlated with zonal Cretaceous standards of the northern Pacific and International Stratigraphic Scale. Generally, local zones correspond to two or more zones of the International Stratigraphic Scale not exceeding the stage volume.

Stratigraphic division and paleobiogeographic reconstructions permit tracing of environmental and biotic development. Initiated in the Late Jurassic, a significant southward expansion of Boreal faunas continued through the Neocomian: East Russia was mostly part of the Boreal zoogeographical realm. At the same time, the late Volginian was marked by Tethyan ammonite migration to as far as 54° N latitude [45].

During the early Neocomian benthic communities were areally dominated by buchias throughout NE Russia; Boreal inoceramids are found occasionally. A *Buchia* scale worked out by Zakharov [46] for Northern Siberia appears to be valid for the early Neocomian of East Russia as a whole. Two zones *Buchia fischeriana* and *B. volgensis* have been recognized in the Berriasian [47]. From work in West Priokhotie, the Berriasian boundary is fixed by the zone *Buchia okensis* [45]. Boreal ammonites similar to Siberian fauna, such as *Neocosmoceras, Chetaites, Praetollia, Surites,* and *Tollia* [46] have been found in the northeast of Russia. During the Berriasian the Far Eastern region was located at the boundary of the Boreal, Tethyan, and Pacific realms. The southernmost Boreal ammonites *Praetollia* and *Tollia* have been discovered nearby 51° N latitude, while the northernmost Tethyan forms at 48° N latitude. The ammonite complex includes *Dalmasiceras, Pseudosubplanites, Beriasella,* and *Spiticeras* and is considered as an analogue of Tethyan communities. The Pacific taxa are represented by *Substeuroceras* and *Paradontoceras* with *Fauriella* from Southern Europe. Tethyan ammonite fauna existed against the background of the Boreal benthos composed of dense populations of *Buchia* [45, 48], a phenomenon most likely due to *Buchia*'s broad temperature tolerance. The southernmost *Buchia* localities are reported from the Nadanhada terrane (NE China) at 45°N paleolatitude. Thus, the ecotone zone (mixture of two ecological communities) was located between 45°N and 50°N during the Berriasian as determined by bivalves [45, 46].

During the Valanginian to early Hauterivian buchias were still dominant and three zones *inflata-keyserlingi*, *sublaevis* and *crassicollis* are distinguished. There are also three ammonite zones in the Valanginian-Hauterivian [47] *Thurmaniceras-Tollia* (s.l.), *Olcostephanus-Dichotomites,* and *Hollisites dichotomus.* The early Hauterivian marks the first appearance of *Coloniceramus* and *Heteropteris* along with the belemnites. Four inoceramid zones are recognized in the Berriasian-early Hauterivian [47]: *Inoceramus (Neocomiceramus) miedae, I. (Anopaea) mandibulaformis, I. (Birostrina) proconcentricus, I. (Neocomiceramus) pochialayneni.*

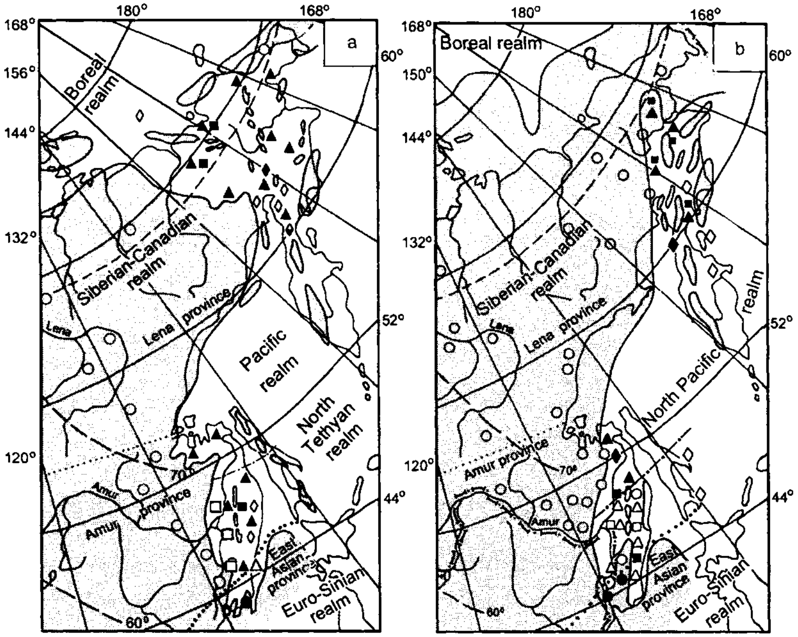
The Boreal fauna dominated in Northeastern Russia during the Valanginian to early Hauterivian, whereas mixed Boreal-Tethyan ammonite faunas are found in some localities of the Far Eastern region (Sikhote-Alin, Primorye): *Tollia*(?), *Kilianella,* and *Homolsomites* (Figure 9a). Based on these data, the northern Tethyan boundary lies between 45°N and 48°N as identified by ammonites, while the southern Boreal boundary is placed at 45°N from bivalves [45, 46].

A viable radiolarian scale for the Boreal province has not been determined yet, but Matsuoka [49] attempts to correlate Cretaceous sequences of the Pacific coastal margin of Russia with well sections of the Pacific bottom. The Berriasian has not been correlated, but the Valanginian-Hauterivian section from the hole 801 coincides with that of the Koryak zone. Matsuoka [49] correlates the zone *Cecrops septemporatus* with the Koryak section based on *Mirifusus chenodes* and *M. mediodilatatus* s.l., and the zone *Dibolachras tytthopora* from *D. tytthopora*. In Koryakia and Kamchatka the Tethyan form *Ristola* and the Boreal *Parvicingula* [50] often occur, explained by tectonic juxtaposition of the sections [28]. Tropical radiolarians do not extend as far as the northern boundary (40°N) of the water mixture [51]. However, because of the water currents a latitudinal zonality in the radiolarian distribution might be misrepresented. At least, mixed Berriasian to Hauterivian radiolarian assemblages appear to be known in East Russia in the interval of 55°N - 65° N [50].

The paleogeography of East Russia changed abruptly in the late Hauterivian as the connection with the Arctic basin was only via the narrow Anyui channel. This change coincided with the Great Boreal Simbirskite transgression in the northern Pacific. This transgression is correlated with global highstand of the second order. From Chukotka to Primorye a marine incursion inhabited by *Inoceramus* of the family Coloniceramidae occurred. Rare Tethyan *Crioceratites* are reported only from southern Primorye. Eastwards, pelagic Tethyan radiolarians migrated to 65° N through the Albian, while the Boreal fauna occurred as far as to 55° N [50].

In the late Neocomian the marine biota changed: *Simbirskites* and *Crioceratites* were the dominant among the ammonites, and *Buchia* were replaced by *Coloniceramus* and *Heteropteris* [47].

The development of the marine basins continued in the Barremian. According to Matsukawa et al. [52], at that time transgression occurred in the south and both northern and south-



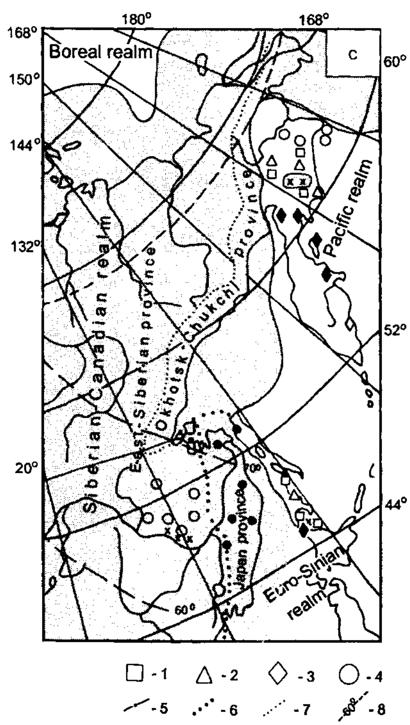


Figure 9. A paleobiogeographical zonation of East Russia in (a) the Valanginian, (b) Albian, (c) - Maastrichtian (from Vinogradov [108]); added and modified from Vishnevskaya and Filatova [36], Sei and Kalacheva [45], Zakharov et al. [46], Pokhialainen [47], Konovalov and Konovalova [48], Vakhrameev [53], Ohana and Kimura [65], Yazykova [69], Markevitch and Bugdaeva [78], and Belyi [84]).

1 - ammonites; 2 - bivalves; 3 - radiolarians: open symbols - Tethyan realm (schemes a, b), Pacific realm (scheme c); shaded symbols - Boreal realm (a, c), North Pacific (b); 4 - paleoflora: open circles - Siberian-Canadian realm; shaded circles - Euro-Sinian realm; 5 - realm boundary from ammonites; 6 - phytogeographic realms boundary; 7 - phytogeographic provinces boundary; 8 - paleolatitudes.

ern water mixed at 45°N. This distribution was in part a result of a deep north-facing embayment [52].

The system of the seas inhabited by *Aucellina* extended along the Pacific coast during the Aptian. The role of the ammonites, in particular *Tropaeum* and inoceramids, continued to be of great stratigraphic importance.

The early Albian saw an expanded transgression and basin differentiation. A chain of marginal and inner seas inhabited by ammonites *Desmoceras* and *Hoplites* and inoceramids stretches along the Asian margin. In the south (Sikhote-Alin, Sakhalin) there were well developed, relatively shallow basins colonized by *Trigonia, Actaeonella,* and *Callista* typical for the Tethyan realm and ammonites *Gastroplites* of the Northern Pacific province [10]. Eastward the marginal seas there were located peri-oceanic basins inhabited by mostly radiolarians.

The mid-Albian in East Russia was marked by a large tectonic reorganization. Since that time a new cycle in marine sedimentation commenced in the Pacific coastal region of Russia, which is related with a considerable renovation in animal communities.

2.2.1.2. Paleoflora

Two Early Cretaceous paleofloristic realms were established in East Russia: the Siberian-Canadian with warm moderate climate and the Euro-Sinian subtropical realm [53]. The former is divided into the Lena and Amur provinces and the latter includes only the East Asian province (Figure 9a). Composition of filices and angiosperms have been chosen as the main criteria for phytogeographic zonation.

The Lena province occupies the northern part of the territory under consideration. The paleoflora of the Lena coal-bearing basin is more thoroughly studied, and three stages, and accordingly three horizons, of the Early Cretaceous paleoflora are established: the Batylykhsky, Ecsenyakhsky and Khatyryksky, which are traceable from the south to the north to a distance of almost 2000 km [54].

The Batylykhsky horizon paleoflora is Neocomian in age and contains more than 160 species. Three floral assemblages are distinguished. The lower assemblage is characterized by *Hausmannia leeana* and *Osmundopsis simplex*, the middle one by *Coniopteris kolymensis* and *C. samylinae*, and the upper assemblage by *Birisia onychioides, Arctopteris heteropinnula, Gleichenites* sp. Among other spore plants *Lycopodites* sp. and *Equisetites rugosus* have been found. Numerous Ginkgoales (*Baiera, Ginkgo, Eretmophyllum, Sphenobaiera,* and *Pseudotorellia*), Czekanowskiales *(Czekanowskia, Leptotoma,* and *Phoenicopsis)* and conifererales (six species of genus *Podozamites*) are predominant among the gymnosperms. Compared to the Late Jurassic paleoflora, the Early Cretaceous paleoflora is enriched by Cycadaceae (*Nilssonia, Ctenis, Pseudoctenis, Jacutiella, Heilungia, Aldania,* and *Doratophyllum*) and Bennettitales(*Nilssoniopteris* and *Pterophyllum*), and Caytoniales (*Sagenopteris lenaensis*) are found in a minor amount. The latter are more characteristic of the Euro-Sinian realm.

The Ecsenyakhsky horizon (70 species) is dated as Aptian. The proportion of major plant groups remained the same. *Birisia onychiodes* is the most typical fern there with frequent occurrences of *Gleichenia.* The diversity of *Cladophlebis* and *Coniopteris, Cycadales* and *Bennettitales* was sharply decreased, and *Heilungia, Aldania* and *Ctenis* disappeared. *Anomozamites arcticus, Nilssonia gigantea, Neozamites verchojanensis, Pterophyllum bulunense* and two new species *Nilssoniopteris* had first appeared. The composition of the coniferous plants shows the emergence of the representatives of the genera *Sequoia* and *Parataxodium* .

The age of the Khatyrksky horizon is determined as early to mid-Albian. The paleoflora of this horizon contains 100 species of plants. Numerous ferns are still dominant: A*diatopteris* (five species are known only from this horizon), *Arctopteris, Birisia* (two typical species), and *Asplenium.* New species of the genus *Coniopteris, Nilssonia, Neozamites* and others appear, and the upper part of the assemblage is marked by rare occurrences of small leaf imprints of Angiospermae.

Thus it can be seen that maximal genus and species diversity of Cycadaceae, Bennettitales, and Czekanowskia is observed in the Neocomian and decrease from the Aptian to Albian.

Compared to the Late Jurassic, increasing generic and species diversity in most plant groups occurred in the Lena province during the Early Cretaceous.

Lower Cretaceous coal-bearing deposits are widely distributed eastwards, in the Zyryansky and Omsukchansky basins where the floral composition is very similar to the coeval floras of the Lena basin [53]. Three horizons of the Lena basin correspond (towards the top) to the Ozhoginsky, Silyapsky and Buor-Kemyussky horizons of the Zyryansky basin. The Buor-Kemyussky horizon is marked by the diversity of the fern *Arctopteris* (five species) and abundance of small-leaved Gymnospermae. This horizon is characterized by a stable taxonomic composition of major plant groups found in different paleogeographical environments: in the continental basins, coastal lowlands, marginal continental volcanic uplifts. Paleofloras similar to those of the Buor-Kemyussky have been found throughout the Siberian-Canadian realm [53]. In Canada and Alaska, where among the continental deposits there occur beds with marine fauna, the deposits containing the Buor-Kemyussky type flora are dated as early to mid-Albian [55].

In the southern Okhotsk-Chukchi volcanic belt (the Ulyinsky zone) the Emanrinsky (mid-Albian) and the Arindsky (late Albian) horizons have been distinguished for the last years [56, 57]. The Emanrinsky horizon is represented by *Cladoflebis arctica, Adiantopteris* sp., *Neozamites* sp., *Sphenobaierabiloba., Sequoia* aff. *minuta, Sequoia* ex gr. *concinna, Dicotylophyllum* sp. The Arindsky horizon is composed of *Gleichenia* sp., *Birisia* aff. *ochotica, Elatocladus smittiana, Cephalotaxopsis heterophyla, Pagiophyllum triangulare, Menispermites* sp. Early to mid-Albian floral complex (twenty-two forms) has also been discovered in the Kuidusunsky zone of the Okhotsk-Chukchi volcanic belt [58].

Two more floral localities belong to the Lena province. A scarce Early Cretaceous floral complex like that of the Lena basin has been described from the South Yakutia basin, which is marked by the presence of thermophilic plants.The southernmost locality of the Aptian to early Albian paleoflora in the Lena province is situated on the northern slope of the Stanovoy range. During the Cretaceous, it represented a large drainage for the Lena and Amur-Zeya basins indicating the boundary between the Lena and Amur provinces [53].

The Amur province in the territory of East Russia covers the Bureya, Toromsky and Uda basins (Figure 9).

Flora of the Bureya basin is the richest, in which two Early Cretaceous complexes, the Soloniysky and Chemchukinsky, are distinguished [57]. The Soloniysky complex (Berriasian-Valanginian) is characterized by ferns *Cyathea, Coniopteris, Hausmannia* and other species, Bennettitales - *Pterophyllum, Anomozamites, Pseudocycas*, Cycadaceae - *Nilssonia, Ctenis, Heilungia*, Ginkgoales - *Ginkgo, Baiera, Sphenobaiera*, rare Czekanowskia, ancient coniferous - *Pityophyllum, Pityospermum, Podozamites*. The palynoflora is dominated by the pollen *Ginkgocycadophytes,* fern and Bryales spore, while the pollen *Classopollis* are not abundant [59].

The typical Hauterivian to early Aptian forms of the Chemchukinsky complex are *Lobifolia novopokrovskii, Sphenopteris lepiskensis, Anthrotaxopsis expansa, Elatocladus manchurica,* which replaced the previous dominants. A sharp reduction in cycadophyte plants indicates that the climate became cooler but at the same time the palynoflora was dominated by various thermophilic Gleicheniidites and Schizoceacea together with wet-loving Licopodium- and Bryales-form species. There was a sharp reduction in the taxonomic diversity among the fern-form and gymnospermous plants. Earlier it has been noted [53] that the composition of the Chemchukinsky complex shows a sharp decrease in species abundance close to that of the East Asia province.

The paleoflora of the Kyndal complex differs greatly from that of the Chemchukinsky. It is composed of the ferns *Ruffordia, Asplenium, Polypodites,* and *Birisia,* and rare Ginkgoceae: *Ginkgo, Elatocladus,* and *Sequoia.* Quite a new element of this flora appears to be Angiospermae represented by leaf imprints *Araliephyllum, Cinnamomoides, Lindera, Celastrophyllum, Dalbergites,* and *Cissites.* The whole flora is determined as early to mid-Albian that was evidenced by rare occurrence of angiospermous small-leaved remains. However, at present, the Kyndal suite including this paleoflora is regarded as Albian-Cenomanian. Together with flora, it contains brackish-water fauna *Trigonioides, Nippononaia* and foraminifers [57].

Another area of the Amur province is the Toromsky basin, from which two floral complexes of different ages have been recorded [60]. The lower complex is dated as Berriasian and resembles the Soloniysky complex, and the upper complex (Albian) has much in common with the Kyndal complex, although the number of species is much greater (about 60 species). The most typical taxa are ferns, coniferous, whereas Bennettitales, Ginkgoceae, Czekanowskia occur but seldom. The Albian is determined by the presence of Angiospermacae similar to those described from the Kyndal complex. A rare Early Cretaceous floral complex is reported from the Udsky basin approximately corresponding to the Berriasian - Barremian [57].

Summarizing the review of the Amur province paleoflora, it should be mentioned that in comparison to the Lena province, which is marked by the floral succession (gradual development) in the Amur province, no intermediate forms between these markedly different floras have been found from the Neocomian to Albian. Vakhrameev [53] suggested that a reorganization took place within the Early Cretaceous and probably in the Hauterivian.

The flora of the Euro-Sinian realm has almost twice the taxonomic diversity of the Siberian-Canadian realm [55]. One of the essential differences between them during the Early Cretaceous is the extinction of Czekanowskia species of the Euro-Sinian realm and the significant presence of Bennettitales and Cheirolipidioceae.

Only Southern Primorye belongs to the East Asian province within the Euro-Sinian realm. Flora and palynocomplexes of the Razdolnensky and Partizansky basins related to this province have been studied by [61, 62, 63, 64, and others].

The floral composition of the East Asian province differs sharply from that of the Amur province [53], although the latter demonstrates some southern elements. First, the Czekanowskia are absent in the floras of Primorye and Ginkgoaceae are relatively scarce. On the other hand, Early Cretaceous floras of Southern Primorye are rich in the ferns *Matonidium, Nathorstia, Ruffordia* and *Weicheselia,* and the Bennettitales *Dictyozamites, Cycadites, Ptilophyllum, Zamites, Zamiophyllum,* and *Williamsonia,* which have not been reported from the Amur province. The thermophilic conifers *Araucariodendron, Podocarpus, Athrotaxites,* and *Brac-*

hyphyllum are abundant. The palynocomplexes are characterized by numerous *Taxodiaceae* and *Gleicheniaceae*, the major coal-forming plants. An early appearance of Angiospermae should be noted, the imprints of which have been reported in the Aptian and pollen as early as the late Hauterivian [13]. Leaves of the Angiospermae *Aralia lucifera* first occur in the Aptian [61]. This information strongly suggests that the floras of Southern Primorye be considered a part of the East Asian province of the Euro-Sinian realm.

In the Partizansky basin there are two floral stages in the Early Cretaceous. A Neocomian-Aptian period was dominated by the ferns *Onychiopsis, Coniopteris* and *Alsophylites*, and the Bennettitales *Dictyozamites,* the Cycadaceae *Nilssonia* and *Ctenis,* and the conifers *Elatides* and *Athrotaxis*. The Late Jurassic floral relicts, *Hausmannia, Klukia,* and *Cyathea*, make up a considerable part of the flora. The palynoflora is characterized by the prevalence of Gymnospermae including numerous *Classopolis* and the first emergence of Schizoceacea, an indicative Cretaceous flora. The late Hauterivian is marked by the occurrence of early angiosperm pollen [13, 59]. *Elatides, Athrotaxopsis, Athrotaxites, Sphenolepidium,* and *Sequoia* are dominant during the early to mid-Albian; *Pityocladus* and *Pityspermum* of the family Pinaceae are also observed. Typical Late Cretaceous ferns *Anemia, Osmunda, Asplenium,* and *Onoclea,* together with individual Angiospermae, are significant [64].

In the Razdolnensky basin Bennettitales, Cycadaceae in association with Caytoniales, high-trunk *Araucariaceae* and *Taxodiaceae* as well as conifers and ferns predominated in the Barremian-Aptian. In the Aptian the first Angiospermae *Pandanophyllum* and *Onoana* appeared [59, 61]. During the mid-Albian, floral composition shows a decrease in variety of Araucariaceae, Taxodiaceae and Cheirolepidiceae as a result in the temperature decrease [61]. Fern-form and coniferous taxa decreased in the palynoflora, while Polypodiaceae, Gleicheniaceae, Schizaeaceae along with Sphagnales and Hepatitae Bryales predominated. The diversity and amount of Angiospermae increased [59].

According to Ohana and Kimura [65], there existed during the Early Cretaceous two types of flora in Japan: the Ryoseki-type flora in the Pacific coast of Japan and the Tetori-type flora in the Sea of Japan coast, and the zone of mixed flora between them. The Ryoseki-type flora resembles that of Primorye and may be truely assigned to the Euro-Sinian realm. Ohana and Kimura consider the Tetori-type flora a part of the Siberian-Canadian realm, while Vakhrameev [53] is inclined to regard it a peculiar plant association of the East Asian province within the Euro-Sinian realm.

2.2.2. Late Cretaceous

2.2.2.1. Paleofauna

A new stage in the evolution of the floral and faunal communities began in the late Albian (Figure 9b) when cardinal tectonic reorganization took place and a chain of the volcanic belts formed creating a complex topography. A cardinal renewal of the animal communities is associated with the period of global sea-level rise. Late Cretaceous seas were inhabited by ammonites, inoceramids, trigoniids and other pelecypods, gastropods, corals, sea-urchins, fish, foraminifers, and radiolarians.

Late Albian regressive marine sediments contain the ammonites *Gastroplites, Neogastroplites, Paragastroplites* and inoceramids, notably *Gnesioceramus* [47]. Marginal seas similar to the late Albian-early Cenomanian Mowry Sea of the U.S. Western Interior [66] occur along the entire Northern Pacific coast and was a fertile environment for abundant foraminifers, especially *Rotalipora,* and accomodated an expansion of the Tethyan radiolarians [50, 67]. In Sakhalin, the fauna is more diverse and includes the ammonites *Cleoniceras, Inoceramus dunveganensis aiensis*, the bivalve *Parvamussium*, the gastropods *Ovactaeonella*, the foraminifers *Saccamina, Hyperammionoides*, and *Orbitolina*, and the radiolarians *Xitus* and *Crolanium*. The Albian/Cenomanian boundary marks the last aucellines, oyster beds, trigoniides, callist coquina in all the seas of East Russia [47].

The mid-Cenomanian transgression covered the whole North Pacific coast and resulted in a major extinction of the biota in the marginal seas. At that time ammonites *Turrilites, Acantoceras*, and *Puzosia* became widespread with new forms *Birostrina concentricus* and *Pergamentia pressulus* among inoceramids. No belemnites are known in the Cenomanian biota [47]. In Primorye the ammonite *Eogunnarites* is known from the early Cenomanian, but inoceramids *I. beringensis, I. ginterensis, I. nipponicus, I. sichotealinensis, I. pressulus* indicate late Albian-early Turonian age. In Sakhalin this group was predominant. It should be pointed out that Late Cretaceous inoceramids in Sakhalin provided the most detailed division (15 zones) and the best correlation as they do not depend on facies. In the Cenomanian, in Sakhalin two zones are established, the zone *Inoceramus crippsi*, and the zone *I. concentricus nipponicus - I. pennatulus*. From the ammonites, along with the *Turrilites* beds there was distinguished a zone *Desmoceras japonicum*, from the foraminifers – the zone *Glomospira corona, Amobaculites gratus* together with radiolarians *Haliomma, Dictyomitra, Lipmanium*, and *Archaeodiciyomitra* [24].

In the early Turonian, the biota of the East Russian seas was testified by inoceramids *Pergamentia reduncus* and *Mytiloides labcatus*. ammonites *Marshalites tumefactus* occurred from the late Cenomanian to mid-Turonian, in Sakhalin there are known beds with *Fagesia* of the early Turonian. After mass extinction at the Cenomanian/Turonian boundary, ammonite diversity remained low, but recorded during the late Turonian [68].

The mid-Turonian-Coniacian transgression was accompanied by abundance and high diversity of ammonites and inoceramids. In northeastern Russia for this time interval there were distinguished four ammonite zones: *Romaniceras ornatissimum, Subprionocyclus* sp., *Forresteria allusudi, Peroniceras* sp., three inoceramid zones: *Inoceramus hobetsensis - I. multiformis, I. memetensis - I. subinvolutus, I. percostatus verus*, and a Coniacian complex of foraminifers, *Archaeglobigerina cretacea* [47]. In Sakhalin the mid-Turonian to Coniacian interval is identified by three ammonite zones: *Scaphites planus, Jimboiceras planulatiformis, Peroniceras* sp., four inoceramid zones: *Inoceramus hobetsensis, I. teshioensis, I. uwajimensis, I. mihoensis*, beds with bivalves *Cuspidaria brevirostris* (Coniacian), two foraminifer zones: *Silicosigmolina futabaensis, Rzehakia subcircularis* and *Planulina rumoiensis, Globotruncana japonica, Trochammina oriovica* and radiolarian complexes - *Spongodiscus impressus* and *Spongurus* sp. [24]. Yazykova [68] analyzed in detail ammonite and inoceramid zones, as well as stages of the organism evolution and suggested her own version to determine a boundary between the Turonian and Coniacian. The appearance of the endemic *Inoceramus uwajimensis* and cosmopolitan *I. (C.) rotundatus* serves a major criterion for Northeastern Russia and the emergence of cosmopolitan organisms *Jimboiceras mihoense* and *I. uwajimensis* for Sakhalin. Moreover, this boundary marks the disappearance of *Subprionocyclus* sp. and *Jimboiceras planulatiforme* and the appearance of *Forresteria alluaudi* and *Peroniceras* sp., two important cosmopolitan taxa [68].

In the Santonian-Campanian, situation was as in the Early Cretaceous when there existed marginal, intracontinental seas and oceanic shelves (radiolarian seas). But in the Late Cretaceous this range was shifted oceanwards. In the early Santonian, marine biota was not much diversed, but great differentiation in biota was noted during the second half of the Santonian (when global regression occurred) and in the Campanian. Various communities were associated with of different types, separated by island arcs.

Inoceramids differ slightly both in the types of the seas and in the marginal seas they are accompanied by ammonites, mostly *Pachidiscus,* whereas in the oceanic shelf they are accompanied by radiolarians. Ammonites that migrated from the Tethys are widespread in many parts of the world, while inoceramids dominated the Santonian-Maastrichtian biota, and appear to be endemic in the Pacific paleobiogeographic realm [47].

In the Northeast during the Santonian-Campanian three ammonite zones are distinguished: *Texanites kawasakii, Anapachydiscus naumanni,* and *Canadoceras compressum,* four inoceramid zones: *Platyceramus undulatoplicatus, Inoceramus naumanni, Schmidticeramus - Pennatoceramus,* and *Inoceramus pilvoensis,* and a foraminifer complex *Globotruncana fornicata - Hedbergella holmdelensis* [47].

In Sakhalin in the Santonian-Campanian three ammonite zones have been revealed: *Anapachydiscus (Neopachydiscus) naumanni - Eupachydiscus haradai,* beds containing *Pachydiscus* aff. *egertoni* (s.l.), *Canadoceras multicostatum*; five inoceramid zones: *I. amakusensis, I. nagaoi, Pennatoceramus orientalis, Schmidticeramus schmidti, I.* aff. *balticus*; three bivalve zones: *Lucina (Myrtea) ezoensis, Anomia subovalis,* and *Nanonavis sachalinensis brevis*; two gastropod zones: *Pseudogaleodea tricarinata* and *Helcion giganteus*; three foraminifer zones: *Nuttallides takayanagii, Globotruncana hanzawae, Quadrimorphina allomorphinoides, Frondicularia striatula, Silicosigmolina compacta,* and *S. californica,* and a Campanian radiolarian complex *Pseudoaulophacus floresensis, Eusyringium livermorensis* [24]. Detailed investigations of the ammonites from Sakhalin show that almost complete extinction of two ammonite families occurred sumultaneously with the Santonian transgression (Figure 10), and only by the end of the Santonian and in the Campanian did new ammonite species begin to appear [69]. A comparison with the zonal scale in Japan reveals that in Japan this crisis was also manifested (Figure 11). Yazykova [69] and researchers from Japan [70, 71] proposed a Santonian/Campanian boundary between the first appearance of *Anapachydiscus (Neopachydiscus) naumanni* and *Inoceramus nagaoi.*

The Campanian is associated with a flourishing of all biota. Large, almost giant inoceramids, ammonites, and gastropods along with the abundance and the great diversity of many organisms indicate that conditions were optimal. The flourishing of the family Pachydiscidae in Sakhalin can serve as an example (Figure 12).

In the late Campanian and in Maastrichtian inoceramids became less abundant, but *Pachydiscus* still predominated among ammonoids. In the late Campanian regression, a considerable ammonite turnover occurred [68] (see Figure 10). In the northeastern Pacific margin coastal region the Maastrichtian is divided into lower and upper stages. The upper Maastrichtian zone is characterized by the ammonites *Pachydiscus neubergicus, Hypophylloceras, Gaudryceras, Neophylloceras, Baculites, Diplomoceras, Didimoceras, Gliptoxoceras,* and *Pseudoxybelloceras,* the inoceramids *Inoceramus (Spiridoceramus) kushiroensis - I. (S.) shikotanensis,* and the foraminifers *Silicosigmolina perplexa* and *Spiroplectamina grzybowskii.*

During the late Maastrichtian, newly formed marine basins in Northeastern Russia represented narrow, extended shallow-water bays, and contained a reduced diversity of late Maastrichtian ammonites (Figure 9c). The *Pachydiscus subcompressus* zone is distinguished there. Early Maastrichtian assemblages of inoceramids replace the specific forms - Koryakia - (the *Korjakia kociubinskii* zone). They disappeared at the Maastrichtian/Paleogene boundary.

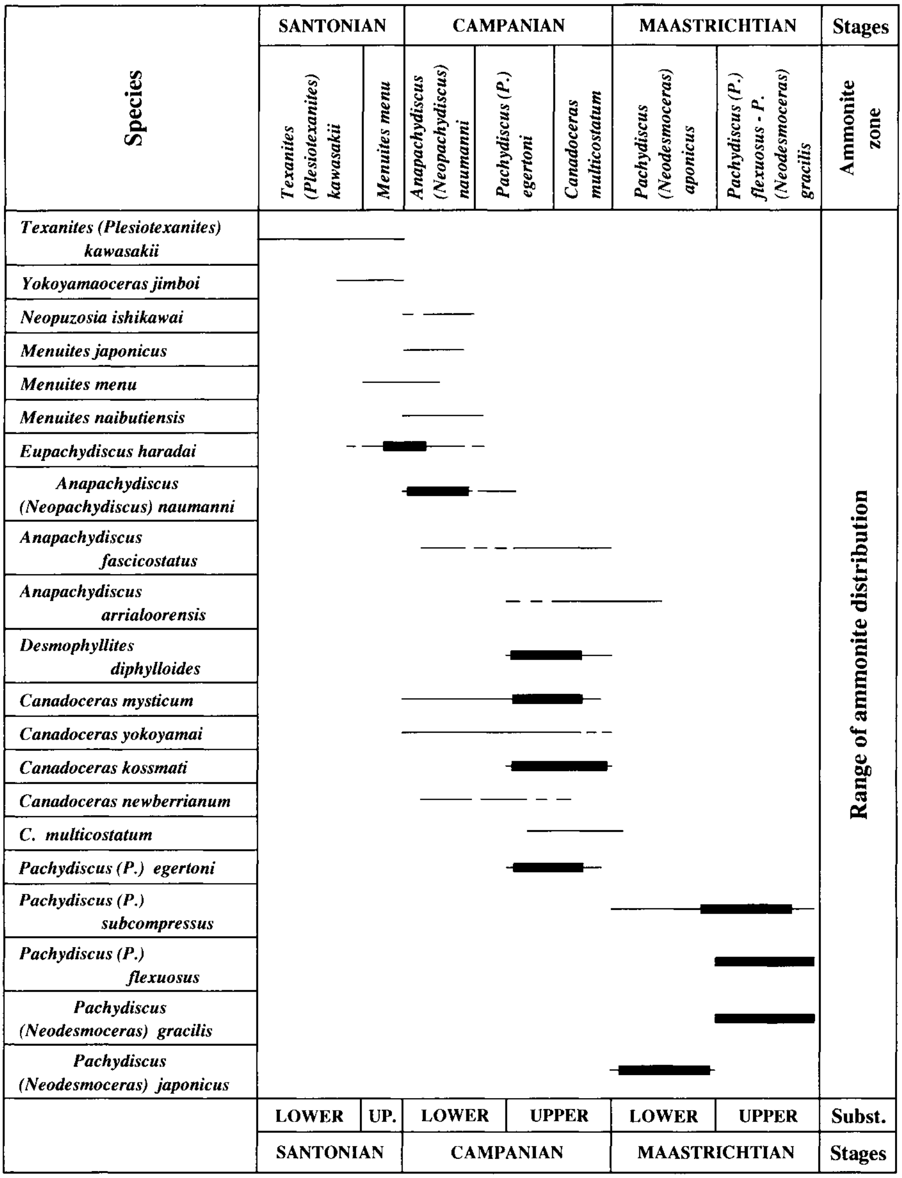


Figure 10. Stratigraphic distribution of the Santonian-Maastrichtian ammonites (superfamilies indicate maximal distribution.

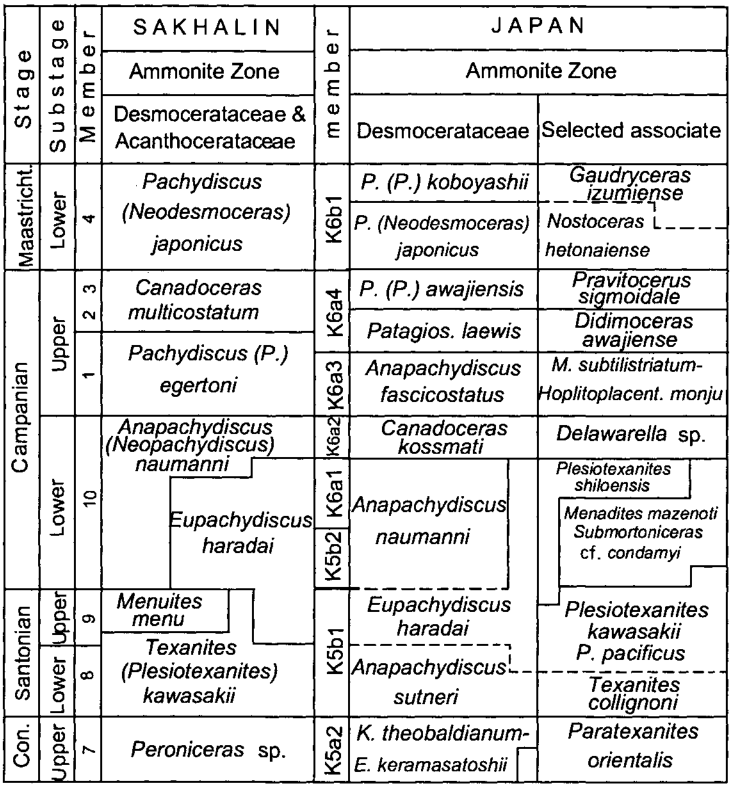


Figure 11. Zonal Santonian-Campanian ammonite divisions in Sakhalin (from Yazykova[69]) and in Japan (from Toshimitsu et al. [70]).

Beds containing the foraminifers *Bulimina kikapoensis* along with the beds of *Globigerina minutula*, make a Maastrichtian to Paleogene transition zone *Rzehakia epigona*. A mass extinction of planktonic foraminifers occurred at the Cretaceous/Paleogene boundary. At the same time the composition of benthic foraminifers did not undergo essential changes. The survival of benthic foraminifers was most likely connected with their ability to utilize organic matter from sediments [72]. A substantial radiolarian extinction (more than fifty species) was confined to the Cretaceous/Paleogene boundary [73]. An abrupt increase in the predominance of *Spumellaria* over *Nassellaria* gives evidence for considerable deepening of the basins [72].

In Sakhalin there is established the *Pachydiscus subcompressus* ammonite zone. Two inoceramid zones are distinguished: *I. shikotanensis* and *I. kushiroensis*; in the northeastern part both of these zones are typical only of the early Maastrichtian. When studying the Makarov section, Salnikova described a bivalve zone *Neilo cuneistrata, Pleurogramatodon splenden* for the Maastrichtian, Poyarkova described a Late Campanian-Maastrichtian gastropod zone *Serrifusus dakotensis*, and Turenko revealed a Maastrichtian foraminifer zone *Haplostiche naibica, Spiroplectammina grzybowskii* [24].

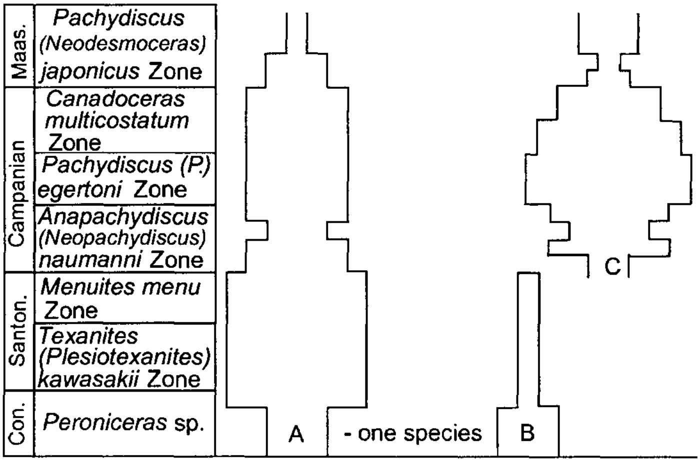


Figure 12. Dynamics of the taxonomic diversity of some Cretaceous families Ammonoidea, showing post crisis recovery of Campanian ammonites.
A - Desmoceratidae; B - Collignoniceratidae; C - Pachydiscidae.

2.2.2.2. Paleoflora

Phytogeographic zonation changes during the Late Cretaceous are minor, as compared to that of the Early Cretaceous. Two realms can be classified in the territory under consideration: the Siberian-Canadian realm, which includes the East Siberian and Okhotsk-Chukchi provinces [53], and the Euro-Sinian realm, which comprises only the Japan province located in southeastern Russia (see Figure 9b, c).

A wide distribution of angiosperms at the end of the Early Cretaceous cardinally changed the floral habitat, and their composition served as a major indicator of Late Cretaceous phytogeographic zonation. So far as angiosperms have already occurred in the Albian floras, floras of that time will also be mentioned in this section.

The East Siberian province occupies the western part of the studied territory. The Agraphen flora seems to be more ancient [74], and is dominated by angiosperms: *Platanus, Pseudoprotophyllum, Araliopsis, Cissites, Celastrophyllum, Maccklintockia,* and *Dalbergites.* Along with the broad-leaved plants, small-leaved plants have been discovered: *Cissites microphylla, Crataegites* cf. *borealis,* and *Celastrophyllum ovale.* Ferns *Asplenium, Birisia,* and *Coniopteris* are also abundant. The coniferous plants are represented by *Araucarites, Brachyphyllum, Cephalotaxopsis, Cryptomerites,* and *Sequoia.* Very rare relicts of the Early Cretaceous flora (*Sphenobaiera* and *Podozamites*) occurred as well. The Agraphen flora is dated by the late Albian to early Cenomanian[53,74].

The Amur-Zeya basin represents the most southwesterly area of the floral distribution in the East Siberian province, from which a small Cenomanian-Turonian complex of the flora *Asplenium, Onichiopsis, Nilssonia, Cephalotaxopsis, Sequoia, Platanus, Trochodendroides, Viburnum,* and *Quereuxia* has been reported. According to Markevitch [75], the Campanian palynoflora of the Amur-Zeya basin was dominated by thermophilic ferns, Cicadophytus and

Taxodiaceae. The amount of the angiosperm pollen, mainly close to Fagaceae, Santalaceae, and Loranthacea, was abundant. The species *Aquilapollenites, Orbiculapollis,* and *Wodehouseia* are very diversified. Species typical for Campanian floras are found among the latter [59].

The Tsagayan flora includes *Ginkgo,* numerous coniferous *Podocarpus, Araucarites, Pinus, Pseudolarix, Sequoia* (rare), *Libocedrus,* and *Cupressinocladus* and the dominant forms - *Metasequoia* and *Taxodium.* Aquatic and semi-aquatic plants *Nelumbo, Potamogeton, Arundo, Phragmites, Limnobiophyllum,* and *Querexia* occur, too. Among the angiosperms predominant were *Trochodendroides, Trochodendrocarpus, Platanus,* and *Tilaephyllum. Myrica, Macklintockia, Menispirmites, Nyssa, Viburnum,* and *Nordenskioldia* occur but occasionally. The first *Celtis* and *Alnus,* which are typical of the Paleogene flora, appeared at that period.

The age of the Tsagayan flora is not clear. It was collected from the Tsagayan suite and is subdivided into three subsuites. In 1969 Bratseva revealed Maastrichtian palinomorphs in the lower and intermediate subsuites. In the early Maastrichtian [59] the thermophilic plants were still dominant [75], but the climate became cooler and drier than in the Campanian. A large plain with small lakes (the Kundur locality) contained dinosaurs Tyrannosauridae, Ornithomidae, Dromaeosauridae, Hadrosauridae, and Nodosauridae, turtles and crocodiles [75, 76]. Plants remains occur sporadically in the early Maastrichtian consisting of *Asplenium dicsonianum, Ginkgo adiantoides, Sequoia reichenbachii, Platanus* cf. *affinis, P.* ex gr. *aceroides* [77].

During the mid-Maastrichtian, the diversity of the thermophilic angiospermous (*Santalaceae* and *Proteaceae)* and gymnospermous plants (Araucariaceae, Podocarpaceae, and Cycadophytes) reduced indicating that the climate was becoming colder.

By the end of the Maastrichtian thermophilic subtropical species became rare. In the floodland forests, Cycadophytus were replaced by Ulmaceae, Myrtaceae, and Taxodiaceae. The broad-leaved, thermophilic plants dominated [59, 78]. In the late Maastrichtian, the role of the Cretaceous relicts remains significant, but broad-leaved angiospermous forms *Limnoliophyllum* and *Trochodendrocarpus* were dominant [77]. The taxonomic composition of the palynoflora reflects these tendencies.

The early Maastrichtian palynoflora of the Bureya basin was slightly different as compared to that of the Amur-Zeya basin. The valley forests were formed by wet-loving plants with the «oculata»-type pollen (*Wodehouseia* and *Orbiculapollis*), and Ulmaceae and Taxodiaceae dominated in the Amur-Zeya basin, probably indicating some differences in the paleoenvironments [59].

Paleobotanic data indicate a deterioration of the climate in the East Siberian province by the end of the Cretaceous. The thermophile subtropical flora gave way to moderately thermophile deciduous plants. The early Maastrichtian warm climate became temperate in the mid-Maastrichtian. Dramatic changes in the taxonomic composition of palynoflora occurred in the upper Maastrichtian, and the floral composition resembled the Tertiary Arctic flora. The changes in palynomorph composition throughout the Maastrichtian conforms with the common ideas of cooling by the end of the Cretaceous. However, the main ecosystem changes began in the mid-Maastrichtian [78].

The Okhotsk-Chukchi province is bounded by the Kolyma river basin from the west. The southerly boundary slightly shifted in the meridional direction due to the climatic changes. At the beginning of the Late Cretaceous it traced through the southern extremity of Sakhalin, while in the Santonian to early Campanian it shifted almost 1000 km northwards.

Late Albian floras are characterized by a specific combination of Early to Late Cretaceous elements almost throughout the Siberian-Canadian realm. They have been reported from the southwestern Okhotsk-Chukchi volcanogenic belt (the Armanian flora), North America and Canada. In North America the late Albian age of these taphofloras is identified by the data on their co-occurrence with marine beds containing *Cleoniceras* sp., *Gastroplites* sp., *Inoceramus anglicus* and *I. caddotensis.*

Angiosperms embrace 32 % of total amount of taxa in the best-studied Armanian flora. Among them predominant are the fern *Birisia* and coniferous of the genera *Podozamites, Cephalotaxopsis* and *Sequoia*. Remains of the typically Early Cretaceous plants known from the Buor-Kemyus-type taphofloras are also present: *Arctopteris kolymensis, Asplenium popovii, Acrostichopteris longipennis, Lobifolia holttumi, Phoenicopsis* ex gr. *angustifolia, Pagiophyllum triangulare, Ranunculicarpus quinquicarpellatus.* Representatives of the genus *Nilssonia* are dominant among the cycadophytes. The taphoflora of the Armanian suite has been marked by the very first occurrence of authentic Platanaceae (*Platanus, Credueria,* and *Protophyllum*) in Northeastern Asia and the representatives of the genera *Populites, Hedera, Menispermites, Hollickia,* and *Trochodendroides;* their leaves are mainly middle-sized, but broad-leaved plants were rare [79, 80]. In the southern part of the Okhotsk-Chukchi belt, above the Arindsky late Albian floristic horizon, the Amkinsky (late Albian-early Cenomanian) horizon lies, in which conifers along with *Metasequoia cuneata, Araukarites* ex gr. *anadyrensis, Trochodendroides* ex gr. *arctica* are predominant [56, 57, 58].

Within the Okhotsk-Chukchi volcanogenic belt, the Armanian-type floras have been reported only southward 64° N and the Chaun-type taphofloras developed in the northern part of the volcanogenic belt [30].

The peculiarity of the Chaun flora is indicated by such plants as *Tchaunia tchaunensis, Tch. Lobifolia, Kolymella raevskii, Cladophlebis grandis, Cl. tchaunensis, Cl. Tchuktchorum,* new species *Ctenis, Elatocladus zheltovskii, Araucarites subacutensis*, and *Trochodendroides microphylla*, and the combination of ancient (*Coniopteris, Birisia, Arctopteris, Ctenis, Heilungia, Ginkgo, Phoenicopsis,* and *Sphenobaiera)* and young elements *(Taxites, Sequoia, Metasequoia, Menispermites,* and *Quereuxia angulata.* Almost all indicative localities of the Chaun flora include *Tchaunia* and *Ctenis.* No other older Cretaceous flora of Northeastern Asia contains the same amount of cycadophytes (some new species of the genus *Ctenis* and *Heilungia*).

Tchaunia is typical almost for all Armanian-type taphofloras, whereas the fern *Birisia ochotica* is an indicative of the Chaun flora. But no *Ctenis* and *Heilungia* are observed in the Armanian flora, and *Nilssonia,* typical of the Armanian flora, are not found in the Chaun flora. The Armanian flora is rich in diverse angiosperms, and the Chaun flora is characterized by its poverty except uncommon in the Armanian flora *Quereuxia angulata,* which is widespread in the Chaun flora. *Cinnamomoides ievlevii* and *Trochodendroides microphylla* occur in the Chaun flora.

For all the differences between the Armanian and Chaun floras, their taphocoenoses are similar. Both in the Armanian and Chaun floras taphocoenosises of the mesophyte (*Tchaunia, Birisia, Coniopteris, Ctenis,* and *Heilungia)* and cenophyte types *(Quereuxia* and *Metasequoia*) usually occur but isolated. Accordingly, both floras reflect a period of the vegetation development, during which mesophyte and cenophyte types vegetation coexisted with each other probably occupying different ecological niches [81].

Besides the northern Okhotsk-Chukchi volcanogenic belt, the Chaun-type flora is also observed on the right bank of the Kolyma river. Thus, in the late Albian a local botanical areal of the Chaun-type flora has been formed.

The emergence and existence of the Chaun flora coincided with a global cooling, the maximum of which was fixed in the late Albian (~100 Ma), and a large structural and relief reconstructions in Northeastern Asia. A wide distribution of the cycadophytes *Ctenis* and *Heilungia* in Northeastern Asia permits a suggestion that the generation of the Chaun local flora might be caused by the local temperature anomaly. It was probably due to the warm water currents that migrated into the eastern Arctic from near the Equator Atlantic via the Mowry Sea (Western Interior Basin) and Cretaceous seas existing *in situ* of the modern Canadian Arctic archipelago.

In the general successive range of the Late Cretaceous floras, the Armanian flora and «Upper Blairmore flora» are followed by the Grebyenka flora in Northeastern Asia, Dunvegan - in Canada and its analogues in the Kolvillsky trough in Alaska [82, 83, 84]. Principal coevalty of the above taphofloras is evidenced by taxonomic composition and co-locality with the marine beds containing *Inoceramus dunveganensis*. As a whole, this flora is interpreted as the Cenomanian though in Northeastern Asia in the stratotypical area of the Grebyenka-type flora distribution, its age is determined as the late Albian to early Turonian [85, 86].

The Grebyenka flora is dominated mostly by angiosperms (~ 37 %, mostly broad-leaved species) and cenotypical conifers (~ 24 %), the diversity of which is much greater as compared to that of the Armanian flora. Ferns make up ~21 %; among them the representative Armanian flora forms of the genera *Birisia, Arctopteris* occur. The Grebyenka flora often contains *Nilssonia*, but *Podozamites* is absent; the remains of *Phoenicopsis* occur extremely rarely except for the lower flora-bearing beds [80, 84].

The Grebyenka taphofloras appear to be the most ancient among the known Late Cretaceous paleofloras, which inhabited the newly formed Penzhin-Anadyr region. The paleoflora existed in rather stable conditions for almost 8 Ma. Like previous floras, the Grebyenka flora and other similar floras were widespread within the Siberian-Canadian paleofloristic realm.

In the Okhotsk-Chukchi volcanic upland, probably at the Albian/Cenomanian boundary, the Chaun flora was replaced by the Arkagalinsky flora [84] that was distributed throughout the territory, while beyond its boundaries it has been observed only in the Arkagalinsky basin. Stratigraphic and palynological data provide evidence for the existence of the Arkagalinsky-type flora untill the end of the Santonian [84].

In the southern Okhotsk-Chukchi belt there were distinguished three floristic horizons corresponding to the stratigraphic range of the Arkagalinsky flora. The Dukchandinsky horizon (Cenomanian) is represented by *Birisia* sp., *Taeniopteris* sp., *Araucarites anadyrensis*, *Menispermites parafavosus*, *Platanus embicola*, *Paraprotophyllum grewiopsoides*, and *Grebenkia* sp. This horizon is correlated with a stage of relative warming [56]. The Uerekansky horizon (early Turonian) is represented by *Arctopteris penzhinensis, Ginkgo uerekanensis*, *Nilssonia alaskana*, *Phoenicopsis* ex gr. *angustifolia*, *Elatocladus smittiana*, *Metasequoia cuneata*, *Trochodendroides* ex gr. *arctica*, *Dalembia vachrameevii*, and *Menispermites* sp. The floral composition indicates that the climate became cooler. Flora similar like that is known from the lower reaches of the Amur river. The Ketandinsky horizon (late Turonian-Coniacian) is represented by *Birisia elisejevii*, *Sequoia* ex gr. *minuta*, *Metasequoia cuneata*, *Trochodendroides* ex gr. *notabilis*, *Menispermites* sp., *Platanus* sp., and *Nelumbites* sp. This flora corresponds to the stage of relative cooling.

The Arkagalinsky-type taphofloras recorded from the Okhotsk-Chukchi volcanogenic belt seemed to be poorer than those of the Chaun floras in ferns, cycadophytes and Ginkgoceae; no *Tchaunia, Birisia ochotica, Conopteris bicrenata, Ctenis,* and *Heilungia,* typical representatives of the mesophyte plants of the Chaun flora have been found. *Phoenicopsis* ex gr. *angustifolia* is abundant in both floras. The amount and diversity of the coniferous plants increases sharply in the Arkagalinsky taphofloras. The angiosperms are more diversified in the Arkagalinsky flora than in the Chaun flora, but their remains are also extremely rare except for *Quereuxia angulata.*

The contrast between the earliest Late Cretaceous floras dominated in the Okhotsk-Chukchi and Penzhin-Anadyr regions appears to be very striking and is manifest in the distribution of the mesophyte relicts and Cenozoic-type angiospermous plants. Thus, the taphofloras of the volcanogenic belt are characterized by the Czekanowskia (first of all *Phoenicopsis*) up to the lower Cenonian boundary inclusively, whereas in the Penzhin-Anadyr region isolated *Phoenicopsis* have been discovered only in the lower upper Albian Grebyenka taphoflora. *Nilssonia* characteristic of both the Grebyenka and most other floras of the first half of the Late Cretaceous appear to be unknown in the Arkagalinsky-type taphofloras.

The abundance of the remains and broad leaves of angiosperms are specific features of the Grebyenka taphofloras throughout their distribution (late Albian to early Turonian), while in the Arkagalinsly-type taphofloras Angiospermae are extremely rare and have small leaves. Moreover, *Trochodendroides,* which is very rare in the Grebyenka taphoflora, appears to be common among the Arkagalinsky taphoflora.

The taxonomic composition and peculiarities of the Chaun and Arkagalinsky floras distribution show that during the cenophyte flora development in the specific paleogeographic environments they were influenced mainly by large structural and topographic transformations and warm sea water currents, and a rather large areal of the local endemic floras was formed in the Siberian-Canadian paleogeographic realm. As conditions of the local thermoanomaly during the global cooling were favourable for the Chaun flora generation, the Arkagalinsky flora, which replaced it, represent typical vegetation adapted to the conditions of the Okhotsk-Chukchi volcanic upland. Belyi following Vakhrameev, distinguishes the so-called «North Pacific refugium» [84].

It is highly probable that the Chaun and Arkagalinsky paleofloras were significant in the formation of Late Cretaceous phytocoenoses in Northeastern Asia.

In the Penzhin-Anadyr region the Late Cretaceous taphofloras are grouped in independent stages [87]. But only in the middle reaches of the Anadyr river close to the Okhotsk-Chukchi volcanic upland and in the Rarytkin range (the Gornaya river basin) that the continental sedimentation continued for a long time. The former was composed of the previously characterized Grebyenka flora (late Albian to early Turonian), and the latter was made up by the Gornaya river flora (mid-Maastrichtian) and that of the Rarytkin (late Maastrichtian-Danian), which replaced it [88]. In other regions, in which the Penzhin, Kaivayamsky and Barykovsky floras have been established, continental flora-bearing deposits occur between thinner marine beds. The above-described floras resemble more or less some extremely local taphofloras recognized from the localities of the Okhotsk-Chukchi volcanogenic belt adjoining the Penzhin-Anadyr region. Attempts to determine taphocoenoses of the volcanogenic belt by the same age as those close to them in taxonomic composition from the Penzhin-Anadyr region [87, 89], led to serious contradictions.

A possible way to eliminate differences is to assume that local taphofloras of the volcanogenic belt originated in independent isolated volcano-tectonic depressions and intermontane basins prior to the similar floras of the Penzhin-Anadyr region. Later, when favourable conditions emerged, young elements of the local floras of the volcanogenic belt migrated into coastal lowlands and islands. They gave rise to local phytocoenoses burials represented by the Penzhin, Kaivayamsky, Barykovsky and other floras. The Barykovsky stage corresponds approximately to the Delokachansky horizon (Campanian) in the southern Okhotsk-Chukchi belt. It is represented by *Anemia* sp., *Hausmannia* sp., *Pterophyllum* sp., *Ginkgo* ex gr. *Adiantoides*, *Metasequoia cuneata*, *Libocedrus* sp., *Trochodendroides* ex gr. *arctica*, *Macklintockia* sp., and *Menispermites* sp. [56].

In the latest Late Cretaceous floras Golovneva discovered significant regularities permitting to resolve the problem concerning the Cretaceous/Paleogene boundary. One of the most complete continental sections of the Koryak upland that contains rich flora is represented by the section of the Rarytkin suite in the Gornaya river basin (the northeastern Rarytkin range). The Rarytkin suite is filled with lacustrine, swamp, river and deltaic deposits of the near-marine lowland and composed of cyclic alternating sandstones, siltstones and coals of about 2000 m thick. To study the floral evolution at the Cretaceous/Paleogene boundary the distribution of 96 species from 124 localities have been investigated, and a range of floral complexes replacing each other in time has been distinguished. Within every complex a total species diversity and an amount of extincted, newly appeared and intergrade species have been accounted. Moreover, the abundance of the species has been estimated. The analysis of changes in diversity within different ecological groups and the study of the dominant species permitted a conclusion concerning Maastrichtian flora replacement by Danian flora in the Koryak upland.

Flora of the Rarytkin suite is subdivided into two floral complexes: the Gornorechensky (late Maastrichtian) and Rarytkin (late Maastrichtian to Danian) being, in their turn, divided into two subcomplexes, respectively (Figure 13). The age of the floral complexes and subcomplexes has been determined from their correlation with other floral complexes from Maastrichtian-Danian sections of the Koryak upland, which include alternating marine and continental deposits containing both floral and faunal remains [90].

The Gornorechensky floral complex is dominated by *Peculnea, Trochodendroides, Celastrinites, Renea, Dyrana, Platanus, Viburnum, Quereuxia, Palaeotrapa,* and *Corylus.* The coniferous species are not abundant and represented by typical Late Cretaceous *Sequoia minuta, Cryptomerites,* and *Taxites,* and younger elements *Metasequoia* and *Glyptostrobus.* A role of the Ginkgo and cycadophytes (*Nilssonia),* which formed probably monodominant communities, is rather considerable in the flora. Ferns are represented by occasional occurrences of *Osmunda.*

The conifers are more abundant in the Rarytkin floral complex as compared to the angiosperms. The flora is dominated by *Corylus,* some species of *Trochodendroides, Metasequoia, Glyptostrobus* and *Microconium. Taxodium, Platanus raynoldsii, Quercus groenlandica, Arthollia, Rarytkina, Celastrinites, Nyssa, Platimelis, Viburnum, Vitis, Quereuxia,* and *Haemanthophyllum* are typical as well. The remains of Ginkgo are rare. Cycadophytes have not been reported. Ferns occur more often than those of the Gornorechensky complex and are represented by *Onoclea* and *Coniopteris.* Considerable changes in species composition and dominants occurred at the Gornorechensky/Rarytkin complexes boundary. These complexes

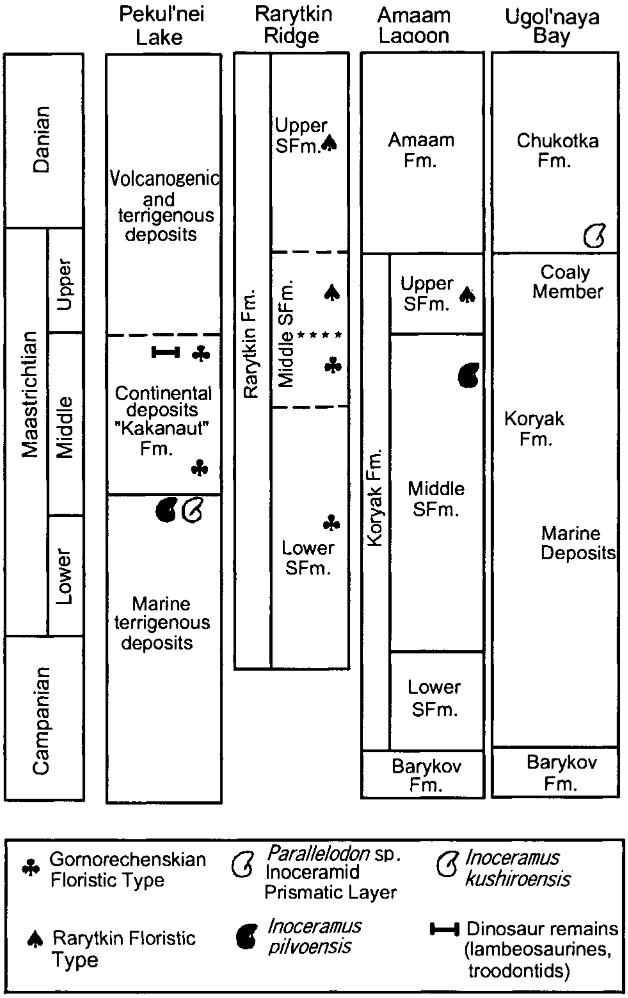


Figure 13. Stratigraphic nomenclature and floristic changes in Late Cretaceous and Early Tertiary rocks of northeastern Russia, Ugolny Bay area (from Nessov and Golovneva [90]).

provide evidence for two different evolutionary stages in the Beringinian paleoflora, which are traceable regionally and are called the Gornorechensky and Rarytkin complexes.

The analysis of the floral development in the Northeast during the Late Cretaceous from the Penzhin [91] up to the Gornorechensky stage reveals that at that time angiosperms developed intensively. There was a 60-80 % turnover in every stage. Coniferous plants, however, experienced minor turnover during the Late Cretaceous and in different floras they are made up of the same genera and even species. The cuticle-epidermic analysis shows that the Late Cretaceous coniferous plants composition were not so monotonous as it has been identified from the data on the morphological study of the sprouts. Nevertheless, many taxa were most likely widespread vertically. The relict elements from Ginkgo, cycadophytes and ferns decreased continuously in the Late Cretaceous floras. The Gornorechensky stage is considered the final stage when typically Late Cretaceous conifers had a considerable part. The Rarytkin stage was characterized by the predominance of the modern genera among the coniferous plants, typical components of the Tertiary flora. At the Gornorechensky and Rarytkin stage the last cycadophytes became extinct, and Ginkgo were not dominant in the communities and became a subordinate component in later floras. Thus, a cardinal reconstruction of gymnosperms occurred at the Gornorechensky/Rarytkin boundary. Typically Late Cretaceous dominants were replaced by the Paleogene ones, and the last mesophyte relicts disappeared. As for the angiosperms, the level of the turnover at that boundary did not exceed that of the previous Late Cretaceous stages, which occurred in the Northeast. More considerable changes in the angiosperms occurred in the late Paleocene, during which the evolutionary rate of the angiosperms increased and representatives of the modern genera from the families Ulmaceae, Betulaceae, Juglandaceae, Fagaceae, and Platanaceae began to have a major part in the flora.

There were 16 common species in the Gornorechensky and Rarytkin floral complexes. They represent the most typical species of the Rarytkin complex that appeared first in the Gornorechensky complex, but became dominant above the complex boundary. The most characteristic taxa of the Gornorechensky complex became extinct abruptly at the upper boundary and did not persist into the overlying deposits. At the early to late Rarytkin complex boundary dominants did not change, but changes in the composition and occurrence frequency of the subordinate species occurred.

The replacement of the Maastrichtian-type flora (the Gornorechensky complex) by the Danian-type (the Rarytkin complex) in the Koryak upland occurred most likely in the mid- to late Maastrichtian rather than at the Cretaceous/Paleogene boundary [90], when in the high latitudes of the northern hemisphere the climate became cooler [92]. Earlier and more considerable cooling in the high latitudes as compared to that in the low latitudes was probably due to the Late Cretaceous/Paleogene increase of the latitudinal thermal gradient [93]. The Maastrichtian/Danian boundary lies within the Rarytkin stage of the floral development and is probably correlated with the events, which took place at the early to late Rarytkin subcomplex boundary. It is concluded that the transition between the Maastrichtian- and Danian-type floras was not synchronous at different latitudes. This appears to contradict catastrophic models for the biota changes at the Cretaceous/Paleogene boundary. It might be supposed that the Danian-type flora has produced in high latitudes during the late Maastrichtian and spread gradually southwards in accordance with the progressive climatic cooling.

The analysis of species diversity change showed an increase in the Rarytkin complex to around 27 % as compared with the Gornorechensky complex. In the late Rarytkin subcomplex a total diversity increases by 37 % as compared with the early Rarytkin subcomplex, but numerous new species of the late Rarytkin subcomplex appeared to be rare. This is at variance with the data obtained by some palynologists, who noted the floral diversity decrease in the Danian as compared with the Maastrichtian. The analysis of the species extinction testifies that all subcomplexes are characterized by approximately equal amount of species extinction, whereas the number of the new species increases in every subcomplex compared to the previous ones. As a whole, the Rarytkin complex includes 70 % of new species. Consequently, the floral change at the Gornorechensky and Rarytkin complex boundary is determined not so much by the extinction as by the appearance of new species. A total amount of the species extinction at this boundary attains 60 %. This is completely correlated with the rate of extinction occurred in other Late Cretaceous floras of Northeastern Asia, which usually makes up 50-80 % [88, 91, 94].

Golovneva [88] concludes that the Danian-type flora formed in the high latitudes since the Maastrichtian and moved continuously southwards as the climate became cooler. Therefore changes in the species composition in the northern floras were not so abrupt, and they were characterized mainly by the change in dominant taxa. In the median latitudes floral changes were caused not so much by evolution as by migration in the sections and appeared to be more abrupt. Krassilov [94, 95] considers the floral complexes change in the Maastrichtian to Danian sections to be associated with the displacement of high latitude plant belts downwards that was due to the fall of temperature. In this case an abrupt change might be caused by species migration rather than by direct extinction and appearance of taxonomic groups.

The southern boundary of the Okhotsk-Chukchi province shifted in the south direction due to the climatic changes. At the beginning of the Late Cretaceous it lay through the southern end of Sakhalin, and by the end of the Santonian it shifted north [53].

At the end of the Albian-Cenomanian, there appeared new coniferous forests with *Sequoia* and *Parataxodium, Cupressinocladus, Protophylocladus,* and *Cycamore-* and *Laurus*-like plants in the undergrowth [96].

As a result of the work carried out by a large group of researchers [97], six Late Cretaceous (including the late Albian) phytostratigraphic horizons have been established for Primorye and particularly for the East Sikhote-Alin volcanic belt [57]. Three of them are related to the Okhotsk-Chukchi province, whereas the other three younger horizons (beginning from late Santonian) belong to the Japan province. The Petrozuevsky horizon (late Albian-early Turonian) is represented by *Gleichenites* sp., *Onychiopsis psilotoides, Asplenium dicksonianum, Cladophlebis frigida, Nilssonia juconensis, Cephalotaxopsis heterophylla, Platanus cuneifolia, Protophyllum* sp., *Viburnum* sp., *Aralicephyllum samargense, Araliopsoides cretacea,* and *Cinnamomorphyllum* sp. The palynomorph assemblages are dominated by Gymnospermaceae: *Taxodiumpollenites hiatus, Inaperturopollenites dubius,* bisaccate close to Pinaceae and Podocarpus. Among the spores the most characteristic are: *Foveosporites cenomanicus, F. canalis, Tricocolpites* sp., *T. variabilis, Retitriletes georgansis, R. vulgeris,* and *Fraxinipollenites constrictus* were abundant among *Angiospermae.* Flora of this horizon shows similarity to the Grebyenka flora [62].

The Arzamazovsky horizon (mid-Turonian-early Coniacian) is represented by *Osmunda asuwensis, Gleichenites nordencladus, Cladophlebis septentrionalis, C. frigida, C. jeliseevii, Ginkgoites pacifica, Elatocladus gracillima, Thuja cretacea, Cryptomeria moiseenkovi, Platanus* sp., *Protophyllum* sp., *Viburnum teutichoense,* and *Physocarpites vachrameevii.* The taxonomic composition of the palynoflora gives evidence for relative cooling [59].

The Kisinsky horizon (late Coniacian-early Santonian) is represented by *Sequoia minuta, S. reicchenbachii, Metasequoia cuneata, Cryptomeria subulata, Cupressinocladus* sp., *Corylus jeliseevii,* and *Viburnum* sp.

The flora of the Razdolnensky and Partizansk basins (Primorye) has been thoroughly studied for the last years. Here at the mid- to late Albian boundary the vegetation experienced a cardinal reorganization.

The late Albian to Cenomanian vegetation is characterized by poor taxonomic composition of ferns and conifers, which prevailed in the Early Cretaceous, by the appearance of *Otozamites* indicating that the climate became arid related to the generation of the Sikhote-Alin volcanic belt, which closed the way to the wet winds. There appeared *Aralia, Sapindopsis, Laurophyllum,* and *Sassafras* along with small-leaved *Platanophyles* [64]. Thermophilic ancient ferns and gymnosperms decreased, being replaced by angiosperms. The palynoflora was dominated by Taxodiceae associated with numerous Bryales, especially Hepatitae and Sphagnales. The Cenomanian complex was dominated by broad-leaved Platanophyle angiosperms, small-leaved plants occur seldom, and ferns are rare [64]. The Cenomanian flora in the Partizansky basin is composed of *Asplenium dicksonianum, Cleichenites* sp., *Dictyozamites* sp., *Coniferites* sp., *Pityophyllum* sp., *Ficus* (?) sp., *Sassafras* sp., *Cercidiphyllum* sp., *Menispermites* (?) sp., *Platanus cuneifolia*, and *Tetracentron* aff. *potomacence* [98].

In Sakhalin during the mid-Albian there existed a marine basin, in which islands were intermittenly emergent. The role of diverse liana-form ferns (Schizaeacea and Polypodiaceae) and Ginkgoaceae, Hirmerellaceae and Gnetales among gymnosperms in the late Albian flora shows a marine cooling influence. In the early Cenomanian the amount of Bryales, Lycopodium-form, Selaginellaceae became more diversified, and the abundance of Taxodiaceae and angiosperms increased. The climate became more humid and cooler [59]. The Turonian flora of South Sakhalin was dominated by ferns, while Taxodiaceae were dominant among the gymnosperms. Angiosperms increased in abundance and diversity in the Coniacian, and numerous Platanaceae and Fagaceae were predominant among them [24, 59]. Broad-leaved Protophyllum as well as Cyatheaceae prevailed among the plant remains [94].

The Santonian saw a global rise in temperature, and province boundaries shifted northwards. Since that time southern Primorye and Sakhalin became a part of the Japan province, which is characterized by the presence of such thermophilic plants as palms of the Euro-Sinian realm.

The Japan province reported from Japan, whereas Santonian to Maastrichtian floras of Sakhalin and Primorye were transitional [53]. In the Santonian-early Campanian, the Sikhote-Alin vegetation was represented by diverse and numerous tropical and subtropical species allowing good correlation with global climate.

The Monastyrsky horizon (late Santonian-early Campanian) is represented by *Asplenium dicksonianum, Sphenopteris stricta, Onoclea* sp., *Cladophlebis borealis, Araucarites longifolia, Glyptostrobus vachrameevii, Menispermites kujiensis, Magnoliaphyllum* aff. *magnificum, Macclinthockia* spp., and *Trochodendroides sachalinensis*. The flora of this horizon with the first findings of thermophile species has much in common with the flora of the Sawayama complex (Japan), the age of which is determined from joint occurrences of *Inoceramus*, and the Jonkiersky (Santonian) complex in Sakhalin [62]. The Kisinsky and Monastyrsky horizons fit in with the palynomorph assemblages of a wide age range (late Coniacian-early Campanian). Dominant among these are thermophile ferns and angiosperms such as *Tricolpites mataurensis, T. lileii, T. gracilis, T. varius, T. sagax, Aquilapollenites asper, A. cruciformis, A. reticulatus, A. trialatus Rouse, Orbiculapollis globosus, O. lucidus*, and *Proteacidites thalmanii*.

The most typical of the Samarga horizon (late Campanian-early Maastrichtian) are the following macroflora remains: *Cephalotaxopsis heterophylla, Taxodium* sp., *Androvettis catenulata, Trochodendroides vassilenkoi, Platanus willjamsii, Protophyllum* sp., *Corylus insignis, C. jeliseevii, Viburnum simile*, and *Menispermites* cf. *nelumbites*. The palynomorph assemblages are dominated by flowering plants mostly producing *unica* and *oculata*-type pollen. The most indicative among them are *Aquilapollenties quadrilobus, A. granulatus, Mancicorpus solidum, M. notabile, Parviprojectus amurensis, Wodehouseia spinata, W. ele-*

gans. Proteacidites crispus, Beaupreaidites oculatus, Erdmanipollis pachyzandroides Krutz., *E. albertensis*, and *Ulmoideipites krempii* .

The Bogopolsky horizon (late Maastrichtian) is illustrated by *Cephalotaxopsis heterophylla, Sequoia* sp., *Androvettis catenulata, Protophyllum* cf. *altaicum, Aryskumia ulmifolia, Aralis* cf. *altaica, Ampelopsis* aff. *acerifolia, Betula lacinibracteosa*, and *Deicampa bicostrata*. The palynomorph assemblages are dominated by angiosperms, mostly *Orbiculapollis lucidus, O. globosus, O. reticulatus*, and *Fibulapollis mirificus*. The most typical are *Aquilapollenites spinulosus, A. minutulis, Tricolpites microscabratus, Triporopollenites plectosus*, and *Triatriopollenites aroboratus*.

According to Krassilov [97], in the East Sikhote-Alin belt vegetation changed radically in composition at the Cretaceous/Paleogene boundary: coniferous forms changed, and birch forests became dominant. The proportion of the pollen similar to the present families increased [59].

During the Santonian, the continental coal-bearing deposits in Sakahlin were dominated by the flora remains, among them species with lobose compound leaves: *Aralephyllum polevoi, Debeya tikhonovichii*, and *Trochodendroides sachalinensis* were dominant [94]. Angiosperms were a significant part at that time, with the prevalence of *Lorantheacae* [59].

In the late Campanian magnoliaphylles together with *Myricarphyllum* as well as the fern *Onoclea glossopteroides* were abundant among the floral remains [94]. The palynoflora was represented by the typical Campanian taxa.

In the mid-Campanian-late Maastrichtian, a considerable regression followed by volcanic activity occurred in Sakahlin and adjoining areas [33]. The climate became warmer, as evidenced by the increased role of thermophyllic communities in the terrestrial floras and considerable diversity of certain groups of molluscs including endemic species against the background of a wide geographical taxa distribution. Specific ecological conditions in the south of Sakhalin were manifest both in a peculiar composition of molluscs (endemic forms, gigantism) and vegetation indicating a wide transition zone between subtropical and warm-moderate zones [24].

Maastrichtian-Danian Avgustovka coal-bearing formation has been described from the Avgustovka river basin in Sakhalin [94] overlying ammonite-bearing marine beds, which contain *Sequoia, Parataxodium, Protophyllum* and *Trochodendroides arctica*. It is in turn overlain unconformably by the Boshnyakovsky suite tuffs containing *Metasequoia* and *Corylites protoinsignis*. Although the Boshnyakovsky flora habit is generally assigned to the Paleocene, many other Cretaceous relicts including *Nilssonia* occur in it that gives evidence to date it by the early Danian. At the boundary between the Avgustovsky and Boshnyakovsky suites major dominants changed abruptly that provides foundation to correlate it with the Cretaceous/Paleogene boundary in marine and continental facies.

The Yurievsky floral complex has been described from the marine deposits containing foraminifers on the Yuri island (the Lesser Kuril Range) [34]. This complex represents a transition zone between the Avgustovka (Maastrichtian) and Boshnyakovsky (Danian) complexes of Sakhalin. The most commonly encountered species are: *Picea* cf. *sonomensis, Pseudolarix* sp., *Sequoia reichenbachii, Cupresssinocladus cretaceous, Androvettia catenulata, Amentotaxus* sp., *Debeya* cf. *pachyderma, Menispermites katiae* sp. nov., *Trochodendroides arctica, Platanus* cf. *heeri, Corylites protoinsignus*, and *Viburniphyllum asperum*. The Senonian dominants among the conifers are still found in this complex, but Danian forms prevailed among the angiosperms. The palynocomplexes demonstrate the same relationships.

A boundary interval (Maastrichtian-Danian) of the Yurievsky floral complex in Sakhalin is absent [94]. At that time uplifting and volcanic activity commenced on the continental margin.

To establish an iridium anomaly sampled representatives from plant-bearing beds at the Cretaceous/Danian boundary on the Yury island have been chosen for a geochemical analysis. The results show a small increase in iridium content as compared to the normal content.

Summarizing the Late Cretaceous biodiversity, it is noteworthy to examine in brief a problem of the dinosaur extinction. There are known six occurrences with dinosaur fossil remains in Russian Far East (see Figure 9c): the Blagoveshchensk, Kundur, Gilchin, Sinegorsk, Kakanaut, and Beringov [90, 99].

Layers containing remains of the hadrosaur *Mandshurosaurus* cf. *amurensis* Riab., the lambeosaur *Amurosaurus riabinini*, teeth of carnivora dinosaurs, as well as the remains of crocodiles and turtles [76, 99] have been extracted from Blagoveshchensk, Kundur, and Gilchin (the middle reaches of the Amur river).

In 1936 Nagao found an incomplete skeleton of *Nipponosaurus sachalinensis* in shallow-marine deposits in southern Sakhalin, within the limits of the town of Sinegorsk.

The northernmost Kakanaut occurrence of dinosaur remains is located at 62°53′N (the Koryak upland), where hadrosaurs, small therapods and bird bones are reported by Nessov and Golovneva [90]. In the Beringov locality plesiosaurs, elasmosaurs, various fish, tracks of crustacens, and abundant plant remains have been discovered.

The palynoflora reflects the character of a Late Cretaceous East Asian flora, and good preservation and abundance of palynomorphs indicates the age of bone-bearing beds as mid-Maastrichtian.

The dinosaur extinction may have been connected with complicated biogeoceonotic crisis rather than catastrophic event [90]. The evidence suggests that Cicadophyte bushes served as main forage for vegetarian dinosaurs and an elimination of this plant may have contributed to the dinosaur extinction by the end of the Cretaceous [100].

Sedimentological investigations and paleogeographic reconstructions along with studies on the bone location and their preservation suggest that dinosaur remains have been buried due to thick sand- and mud-streams that originated from volcanic and seismic activity [75, 76, 90].

2.3. Paleoclimatic environment

This section is based on both original investigations of the IGCP 245 and 350 project participants, and the results of intradisciplinary research initiated in 1993 on the programme «Warm Biosphere» by the Russian Academy of Sciences with special reference to the Cretaceous [5, 101, 102, 103].

Rocks of nonmarine origin (coals, kaolinic crusts of weathering, red beds) and the distribution of various plants and climatically sensitive animals are the attributes we recognize for climatic determinations. Terrestrial plants are regarded to be of most paramount importance. Animal remains are less reliable due to their potential ability to hibernate or migrate. Average annual temperatures obtained from $^{16}O/^{18}O$ or Ca/Mg ratios in shell valves mostly in northern Siberia and the Pacific sections [104].

2.3.1. Early Cretaceous

During the Early Cretaceous, two latitudinal humid belts are distinguished by the predominance of gray terrigenous sedimentation in the abundant coal basins: these we call the northern circum-Arctic coal-bearing, and mid-latitude coal-bearing belts.

The northern circum-Arctic coal-bearing belt comprised regions of Laurasia north of 57°-60° N and had a global distribution during the Berriasian-Barremian [103]. In NE Asia, the Lena, Zyryansky and some smaller Pegtymelsky, Anyui, North Priokhotie, Omsukchan and Taigonos coal fields are located within this belt.

The mid-latitude coal-bearing belt southern boundary lies approximately at 30° N, but somewhat further south in eastern China. In East Russia coals are dominant with occurring kaolinites and bauxites in other parts of the belt [103]. During the early Neocomian the South Yakutiya, Udsk, Amur-Zeya and Bureya coal fields formed within the belt. In the Hauterivian-Barremian an intense coal accumulation also initiated in the Partizansk and Razdolnensky basins (Primorye).

The Cretaceous is regarded as a non-glacial period. Nevertheless, isolated beds occur in some Lower Cretaceous seasonal glacial-marine deposits at high latitudes [105, 106]. In northern Eurasia the average annual temperature did not exceed 15° C during the Neocomian. Seasonal prevalence of climate is confirmed by paleobotanical investigations[101].

Phytogeography shows a more detailed pattern of the climatic zonality during the Cretaceous. Early Cretaceous vegetation of the Siberian-Canadian realm is related to that of the Late Jurassic. Its composition (see Section 2.2.1.2.) suggests a warm moderate climate [53].

Analysis of Early Cretaceous vegetation and palynomorph assemblages shows that the climate was moderate in the Lena province and warm in the Amur province. The boundary between them is drawn at 50° N. The East Asian province within the Euro-Sinian realm was marked by a warm subtropic climate in the Early Cretaceous. Near the Northern Pole the climate was probably cold-moderate [101]. A minor cooling trend coincided with the Valanginian-Hauterivian regression, and a warming trend in South Primorye occurred in the Aptian [96], which is supported by oxygen isotope ratios of the benthic foraminiferal shells [104]. The mean yearly temperature decreased by 5° C in East Russia during the Albian [107]. However, oxygen isotope data from nannoplankton in the northern Pacific indicate a slight warming [104]. A short-term warming has been established in southern Primorye at the end of the Albian as evidenced by palynoflora studies [59] and the composition of the Chaun flora in NE Russia where the warming is considered to have been influenced by warm water currents [84].

A latitudinal climatic zonality was dominant during the Early Cretaceous, and a longitudinal zonality was manifest only by more humid climate along the Pacific coast.

2.3.2. Late Cretaceous

During the Late Cretaceous the latitudinal climatic zonality prevailed on a global scale [101].

In northern Eurasia there existed a high-latitudinal moderate humid belt between latitudes 60° and 85° [101], in which gray terrigenous coal-bearing deposits accumulated. The belt is characterized by seasonal changes in climate, development of forests and swamps, and annual temperatures attained 10-14° C. A large light and related temperature seasonality conditioned by the alteration of polar days and nights appeared to be the reason of frequent climatic fluctuations within the belt, changing from warm-moderate to cold-moderate climate. This is evidenced by faunal and floral composition [53, 55, 95, 107]. The summer abundance of daylight for 3-4 months created a moderately warm and humid climate enabling rapid vegetation growth. The quick transition to dark winter conditions reduced the processes of biological and chemical destruction of the plants, thus favoring their preservation and burial that resulted in intense coal accumulation [101].

The northern mid-latitudinal warm humid belt lay between approximately 30° and 60° N [101]. Gray terrigenous coal-bearing complexes developed within the belt, except southern Primorye, which is famous for Late Cretaceous red beds. The whole belt was dominated by evergreen humid forests, mostly coniferous, and the mean yearly temperatures varied by 13° to 20° C.

Side by side with a global zonality, a sublongitudinal zonality produced in the late Albian. The growth of volcanic ridges (up to 3000 m) in the East Asian volcanic belt created a local perturbation of this zonality and limited access of humid marine winds to the continental regions of East Russia [53].

The development of dry summer to the west of the volcanic belt led to the extinction of plants that require humidity such as ferns, Czekanovskii, most *Ginkgo*, *Nilssonia*, and Keithonievii. The extinction of Bennettitales was most likely reasoned by low winter temperatures [53].

On the basis of the vegetation composition, a relative late Cenomanian to Turonian warming and Maastrichtian cooling is noted in the East Siberian province [59]. Within the volcanic upland of the Okhotsk-Chukchi province a relative warm environment is reconstructed in the Cenomanian and Campanian, but a relative cooling was recognized in the Turonian and Maastrichtian [56].

Paleoclimatic reconstructions have been made for the Late Cretaceous of the Pacific coast of northeastern Asia during the last years compared with those of Alaska [55]. In the Anadyr-Kamchatka subregion (see Section 2.2.2.2.), a relatively warm climate is reconstructed in the late Albian-Cenomanian, a considerable cooling in the Turonian and Coniacian (with a maximum in the Turonian), warming in the Santonian and early Campanian, and cooling by the end of the Cretaceous (see Figure14).

The mean yearly temperatures varied by 6°-8° C in the Turonian and late Maastrichtian, and by 11°-13° C during the Cenomanian and Santonian-early Campanian optima [55]. Warm water oceanic currents from the equator to poles appear to have played a substantial role in the heat distribution of the Earth during the Late Cretaceous. The Arctic basin was rather warm throughout the Cretaceous that determined non-glacial type of the climatic zonality [101].

The humid climate of the Pacific coast is evidenced by the abundance of coal seams alternating with marine sediments in the continental deposits. A warm and humid climate of the coast is manifest by the presence of Bennettitales, *Nilssonia*, the abundance and diversity of which increased southward, and by the aquatic angiospermous *Quereuxia* that grew in the coastal areas of the lakes [53].

In the Japan province of the Euro-Sinian realm a subtropical climate with warming in the Santonian-early Campanian and cooling in the Maastrichtian is confirmed by an abrupt elimination of the pollen amount *Classopolis* [24, 53].

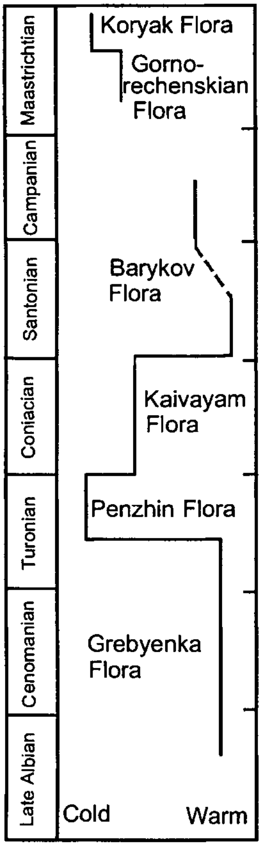


Figure 14. Reconstruction of paleotemperature changes in the Anadyr-Koryak subregion based on paleoflora (from Herman [87]).

3. CONCLUSIONS

Multi-disciplinary studies and stratigraphic correlation carried out by a large group of researchers permitted identification of the synchroneity and interaction of various geological and biological events and reworked the consecutive changes in climatic, biological and physical-geographical environments that occurred in East Russia during the Cretaceous. Here existed an active continental margin, within which sea gradually retreated eastwards.

During the Early Cretaceous the climate varied from cold moderate in the pole to warm humid moderate in the circum-Arctic latitudes, warm in the mid-latitudes, and warm subtropical in southern Primorye. Terrigenous sedimentation prevailed throughout the territory, but the distribution of volcanogenic rocks in both marine (island arcs) and continental conditions was areally limited.

On the continent at that time terrigenous rocks accumulated in fresh-water lakes and river valleys separated by mountain ridges. The lake deposits contained fresh-water fauna (ostracods, conchostracans, fishes). The environment of the huge marginal plains of the Lena river basin was rather specific with substantial coal accumulations in the marshes that indicate that the climate was humid. The vegetation composition indicates that the territory was covered by thick coniferous forests, under which herbaceous ferns, equiseta, certain Cicadaceae grew up. The seasonality of the climate is identified by tree rings. The mean yearly temperature did not exceed 15° C. Northeastern Asia was characterized by a winter short-term fall temperature below O° C. The average temperature of near-surface marine waters abutting Siberia was 10°-12° C.

In the marginal zone there existed deltaic, beach, and lagoon environments, in which mixed marine and fresh-water fauna occurred being replaced eastwardly by shallow sea water environment. During the Neocomian, seas of the Pacific realm were freely connected with the seas of the Boreal and Tethys realms as evidenced by common fauna of buchias and ammonites. To the east deep-water seas were filled with volcanogenic-cherty and clayey sediments and turbidites, and were inhabited by radiolarians. By the latest Cretaceous the Anyui strait, which linked the Boreal and Tethyan regions, closed causing the linkage between them to cease.

The mid-Albian was marked by a considerable tectonic reorganization in the region resulting in substantial changes in physical-geographical, biological and climatic environments. At that time along the continental margin a giant East Asian volcanic belt (up to 3000 m) began to form, thus creating a sublongitudinal climatic zonation against the background of the latitudinal one, which in general inherited that of the Early Cretaceous. It should only be noted that during a Santonian-Campanian warming a latitudinal belt boundary was located almost 1000 km to the north.

In the Late Cretaceous there existed a high-latitudinal moderate humid belt between the latitudes 60° and 85°. It was characterized by ruggedness of the relief, terrigenous sedimentation, seasonal climatic fluctuations, the development of coniferous broad-leaved deciduous forests, and swamps; the average annual temperature was 10°-14° C. The climate within the belt varied from cold moderate to warm moderate that is identified by the vegetation composition, among which angiosperms were dominant.

In the interval between latitudes 30° and 60° there existed a mid-latitudinal warm humid belt. Here in the conditions of the rugged mountainous relief, mostly gray terrigenous coal-bearing horizons accumulated in the basins with the exception of the southern Primorye where red beds were deposited during the late Albian to Cenomanian. An average annual temperature was 13°-20° C. In the Santonian-Campaniana sublatitudinal climate was in the south of the belt (in Primorye) as identified by floral composition. The boundary between the phytogeographic provinces shifted almost 1000 km north.

While discussing the role of the sublongitudinal zonality, it should be mentioned that in general the climate outside the volcanic range was drier and colder. The maritime plain, which existed in front of the volcanic range, was repeatedly submerged in a cycle of alternating shallow-water, deltaic and continental environments. The latter was characterized by an intensive coal accumulation. As earlier, terrigenous sedimentation was dominant.

The active continental margin with typical environmental change of marginal seas - island arcs - facies of the open sea remained untill the end of the Cretaceous. During the Late Cretaceous an essential rejunevation of the fauna occurred. The fauna in the marginal seas was dominated by rapidly evolving inoceramids and ammonites, whereas the open seas linked with the ocean were inhabited by inoceramids, radiolarians and foraminifers. Faunal assemblages at the family level were repeatedly replaced during the Late Cretaceous. Evolution of major representatives of the Late Cretaceous biota of ammonites and inoceramids followed the same laws but somewhat asynchronously. At the Cretaceous/Cenozoic boundary both ammonites and inoceramids became extinct. Radiolarians underwent a mass extinction except some groups.

As for the vegetation, a cardinal reconstruction of ecosystems commenced in the mid-Maastrichtian associated with cooling that in high latitudes began prior to that in the low latitudes. The data on the catastrophic events at the Cretaceous/Paleogene boundary have not been confirmed.

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Cretaceous System in Mongolia and its depositional environments

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The Cretaceous sedimentary basins in Mongolia are grouped into three major depositional provinces (Altai, Khangai-Khentei and Gobi) with ten basins (basins of Great Lakes, Valley of Lakes, Arkhangai, Onon, Choir-Nyalga, Choibalsan, Transaltai Gobi, Umnogobi, Dornogobi and Tamsag), which are all filled by continental deposits with abundant vertebrate and invertebrate fossils as well as plant remains. Most of their stratigraphical, paleontological and sedimentological features are summarized, giving regional correlation of the strata and taking account of environmental characteristics.

1. INTRODUCTION

Nonmarine Cretaceous of Mongolia is unique in many respects, and contains a wealth of paleontological remains of palynomorphs, many different groups of limnic invertebrates and vertebrates. The Cretaceous facies reflects complicated systems of nonmarine deposits.

Since the pioneering expeditions of the American Museum of Natural History in the 1920's, the Gobi Desert has been famous for yielding the first nest of dinosaur eggs, first known dinosaurian hatchings and the oldest placental mammals. Subsequently, paleontological field activities were undertaken by the Mongolian Paleontological Expeditions of the USSR Academy of Sciences (1946, 1948-1949), the Polish-Mongolian Joint Expeditions (1963-1965, 1967-1971) and the Mongolian-Russian Joint Paleontological Expedition (since 1969). The American Museum of Natural History sponsored expeditions to the Mongolian badlands in 1994 resulting in significant and prolific paleontological findings from a new area at Ukhaa Tolgod with abundant articulated skeletons of Late Cretaceous dinosaurs, lizards, and mammals. The Mongolia-Japan Joint Paleontological Expedition sponsored by Hayashibara Museum of Natural Sciences collected dinosaurs and associated fossils from Upper Cretaceous localities, including the dinosaur footprints locality Shar Tsav. The Chinese-Japanese-Mongolian Expeditions sampled Jurassic to Upper Cretaceous strata in South Mongolia during the summers of 1995-1998.

The Mongolian Geological Survey parties have made systematical geological investigation in all areas and compiled a series of 1:200,000 geological maps. In addition, many investigations on specific topics in stratigraphy, regional structure and paleontology have been made in Mongolia by Mongolian geologists. Most of the basic knowledge on the Cretaceous stratigraphy and fossils are written in Russian and Mongolian, but during the last several years the publications in English are increasing.

This paper summarizes published data on the Cretaceous System in Mongolia focusing on the stratigraphy of Cretaceous lithostratigraphic units and their faunal assemblages with special reference to the environmental aspects.

2. OUTLINE OF THE CRETACEOUS SYSTEM

The Cretaceous sedimentary basins in Mongolia are grouped into the three major depositional provinces (Figure 1). The Altai province (A) consists of Basin of Great Lakes (I) and Valley of Lakes (II). The Khangai-Khentei province (B) includes the Arkhangai (III), Onon (IV), Choir-Nyalga (V) and Choibalsan (VI) basins. The Gobi province (C) includes basins of Transaltai Gobi (VII), basins of Umnogobi (VIII) and Dornogobi (IX), and the Tamtsag depression (X). These provinces record different sedimentation and tectonics.

Nonmarine Cretaceous rocks are widespread throughout Mongolia and are represented by continental deposits, plutonic and volcanic rocks [1]. The stratigraphical subdivisions of the Cretaceous strata in various regions are shown in Table 1.

2.1. Altai province

2.1.1. Basin of Great Lakes

Widely distributed Lower Cretaceous sediments are divided into the Gurvanereen and Zereg Formations [2] distributed in the Sangiin dalai nuur, Zereg and Ikhes nuur depressions. In the Khar us nuur and Khuisiin gobi depressions, only the Zereg Formation is exposed.

The Gurvanereen Formation (120-500 m) is characterized by the alternation of sandstone-slate and thin-laminated siltstone, mudstone, and marl, containing the ostracods *Daurina mongolica, Cypridea vitimensis, C. trita, C. zagustaica, C. trita, Mongolionella palmosa, M. martini, Darwinula contracta, Torininia* sp., *Rhinocypris* sp., *R. potanini*; mollusca *Valvata subpiscinalia*; fish *Stichopterus popovi, Gurvanichthys mongolicus*; insects *Ostracindusia bais-*

Table 1

Stratigraphic correlation of Cretaceous sediments between regions

Time Scale	Region. Subdiv.	(A) Altai			(B) Khangai - Khentei					(C) Gobi						
		Great Lakes (I)	Valley of Lakes (II)		Arkhangai (III)	Onon (IV)	Choir - Nyalga (V)		Choibalsan (VI)	Transaltai Gobi (VII)	Umnogobi (Southern Gobi) (VIII)			Domogobi (Eastern Gobi) (IX)		Tamtsag (X)
Age	Formation	Khosbayar [2]	Shuvalov [3]	Sinitsa [5]	Khosbayar [6]	Khosbayar [6]	Khosbayar [6]	Matsukawa et al. [7]	Khosbayar [6]	Shuvalov [3]	Shuvalov [3]	Lopatin [17]	Khand & Badamgarav [18]	Martinson et al. [20]	Shuvalov [3]	Khosbayar [6]
Up. Paleoc.	Khashaat									Naranbulag						
Maastricht.	Nemegt									Nemegt	Nemegt			Javkhlant	Nemegt	
Campanian	Baruungoyot									Baruungoyot	Djadokhta		Bayanzag		Baruungoyot	
Coniacian-Santonian	Djadokhta															
Cenomanian Turonian	Bayanshiree									Bayanshiree	Bayanshiree		Bayanshiree	Bayanshiree	Bayanshiree	Bayanshiree
Albian-Cenoman.	Baruunbayan		Baruunbayan										Baruunbayan	Sainshand		Sainshand
Aptian-Albian	Khukhteeg	Zereg	Khulsangol	Khulsangol Boontsagaan Series: Kholboot Beds	Shankh	Bayankhaan	Uvdugkhudag Choir		Aduunchuluu	Baruunbayan Dushuul	Baruunbayan Ulaandel	Sainshand Khukhteeg	Khukhteeg	Khukhteeg	Baruunbayan Khukhteeg	Zuunbayan
Hauterivian Barremian	Shinekhudag		Andaikhudag	Boontsagaan Series: Khuriit Beds Shartolgoi Beds	Battsengel	Zaanshiree	Nyalga	Shinekhudag	Gurvanzagal			Shinekhudag	Shinekhudag	Shinekhudag	Shinekhudag	
Valanginian-Berriasian	Tsagaantsav	Gurvanereen	Undurukhaa	Undurkhin	Arkhangai	Khongork-hairkhan	Kherlen	Tsagaantsav	Sumiinnuur Bayantumen Ochirkhureet		Tsagaangol	Manlai Mogoit	Manlai Tsagaantsav	Tsagaantsav	Tsagaantsav	Tsagaantsav
Tithonian-Kimmeridgian	Shariliin	Ikhesnuur	Toromkhon	Dundarga-lant Series: Shandrol Beds Andakhudag Beds	Jargalant	Seruungol	Dorgot	Sharliin					Shariliin	Shariliin	Shariliin	

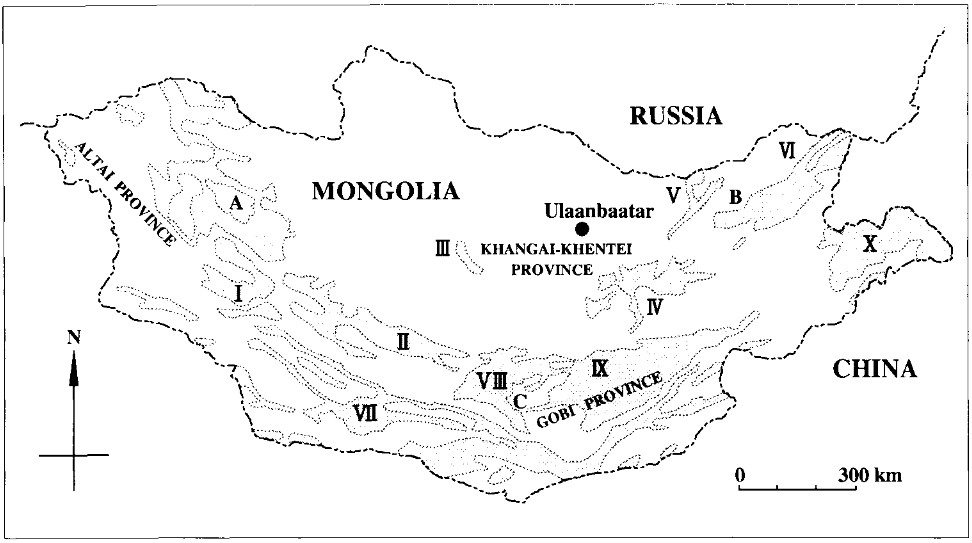

Figure 1. Distribution of Cretaceous sedimentary basins (shaded) in Mogolia.
A: Altai province. I: Basin of Great Lakes, II: Valley of Lakes; B: Khangai-Khentei province. III: Arkhangai, IV: Onon, V: Choir-Nyalga, VI: Choibalsan; C: Gobi province. VII: Basins of Transaltai Gobi, VIII: Basins of Umnogobi (Southern Gobi), IX: Basins of Dornogobi (Eastern Gobi), X: Tamtsag depression.

sica, Corixonecta hosbayari, Secrindusia translucens; plant fossils such as *Eciusetum sp., Otozamites* sp., *Pterophyllum* cf. *acutilobum, Brachyphyllum densiramosum, B. sulcatum, Pseudolarix erensis, Pityolepis* sp., *Problematospermum* sp., *Tyhpaera fusiformis,* and the angiosperm fruits *Gurvanella dictyoptera* and *Erenia stenoptera.* The Gurvanereen Formation conformably overlies Upper Jurassic beds and conformably underlies the Zereg Formation.

The Zereg Formation (150-450 m), which forms the uppermost part of the Lower Cretaceous succession in Western Mongolia, is composed of reddish-brown, gray and yellow sandstone, siltstone, clay, calcareous shale, and marl layers, and yields the ostracods such as *Darwinula* ex gr. *oblonga, D. contracta, Cypridea zagustaica, C. gurvanensis, C. prognata, C. vitimensis, Rhinocypris* ex gr. *jurassica, R. barunbainensis, R.* ex gr. *tugurigensis, R. potanini*; insects *Ostracindusia modesta, O. baissica, O. onusta, O. conchifera, Pelindusia ostracifera, P. minae,* and *Cristocorixa gurvanica,* and mollusca from genera *Unio, Leptesthes, Limnocyrena, Valvata* and *Bithynia.* The Zereg Formation is covered unconformably by the Cenozoic deposits.

2.1.2. Valley of Lakes

Lower Cretaceous deposits are well developed and widespread in the Toromkhon, Jargalant, Bakhar, Kholboot, Khukhtolgoi, Khulsyn gol, and Andai khudag depressions of the Valley of Lakes, and are divided into the Undurukhaa, Andaikhudag, Khulsyngol and Baruunbayan Formations [3].

The Undurukhaa Formation (165-1000 m) rests on the Upper Jurassic Toromkhon Formation conformably and consists of volcanic sediments and sediments of gray, gray-green marls, limestones, and sandstones with conglomerate and gravelite. The fauna is dominated by molluscs such as *Arguniella ovalis, A. asistica, A. quadrata, Lunocyrena resta, L. globosa, L. marginata,* and *Viviparus robustus*; ostracods *Cypridea trita, Mongolianella palmosa, Darwinula contracta*; and conchostracans *Pseudestherica erdeniulensis* and *P. dahuricus.*

The Andaikhudag (= Ondai Sayr by Berkey and Morris [4]) Formation (150-700 m) overlies conformably beds of the Undurukhaa Formation and consists of dark gray, black bituminous paper shales, siltstone and aleurite. Rhythmic alternations of sandstone, siltstone, clay and paper shale are observed. Remains of molluscs such as *Limnocyrena submarginata* and *Zaptychius lacustris*; ostracods such as *Cypridea* aff. *koskulensis, Darwinula* cf. *barabinskiesis* and *D. contracta*; conchostracans such as *Estheria (Bairdestheria) middendorfii*; fish such as *Lycoptera middendorfii*; insects such as *Idnasia reisi* and *Ephereropsis trisetali*; and the fossil plant *Nilssoniopteris* cf. *barabinskiesis.* This fossil assemblage suggests the Hauterivian-Barremian age.

The Khulsangol Formation (600-700 m) overlies the Andaikhudag Formation with erosional and stratigraphic disconformity. It is represented by sandstone, conglomerate, aleurite and clay with intercalated beds of coaly shale and marl. Aptian-Albian molluscs such as *Limnocyrena anderssoni, L. tani, Campeloma claviithiformis,* and *Viviparus onogoensis*; and the plant fossils *Brachyphyllum japonicum* and *Parataxodium jacutensis* have been identified from this formation.

The Baruunbayan Formation (180-200 m) with angular unconformity overlies the Khulsangol Formation and is characterized by alternating beds of red conglomerate, sandstone and aleurite.

On the base of the bio- and lithostratigraphical studies the Andaikhudag Formation in the Toromkhon, Jargalant, Bakhar, Kholboot, Khukhtolgoi, Khulsyn gol and Andai khudag depressions were referred to the Upper Jurassic. The Lower Cretaceous Undurkhin (= Undurukhaa) and Boontsagaan Series as well as the Khulsangol Formation were newly established [5]. The age of the Undurkhin Formation is the same as established by previous authors and the Tsagaantsavian. The age of the Boontsagaan Series is regarded as the Shinekhudagian and Khukhteegian.

2.2. Khangai-Khentei province

2.2.1. Arkhangai basin

The Cretaceous rocks in this basin are represented by the Arkhangai, Bartsengel and Shankh Formations that are distributed in the Ulaankhujir, Bayanduurkhi, Bayantsagaan and Khotont depressions.

The Arkhangai Formation (250-300 m) conformably overlies the Upper Jurassic Jarantai Formation and consists of white gray sandstone and siltstone with intercalations of coaly shale and coal that contain the insects *Ephemeropsis, Coptoclava* and *Hemerobcopus*, and the plant fossils *Phoenicopsis angustifolia, Pityophyllum* sp. and *Todites* sp. The age is early Neocomian. The Battsengel Formation consists of basalt and trachyandesite. On the basis of stratigraphic position they were referred to the late Neocomian age. The description of the Shankh Formation has not been published.

2.2.2. Onon basin

Cretaceous sediments occur in the Onon, Batshireet, Galttai and Tsagaan nuur depressions of the Onon basin, and are subdivided into the Khongorkhairkhan, Zaanshiree and Bayankhaan Formations.

The Khongorkhairkhan Formation (150-200 m) is well developed in the Dadal sum area, Onon and Urt River basins, and rests conformably on the Upper Jurassic Seruungol Formation. It is composed of clay, thin-bedded siltstone, clay shale, and shale sometimes with intercalations of conglomerate and sandstone that contain the insect *Chironomaptera* sp. and conchostracans *Estheria grandis, Bairdestheria* sp. and *Pseudoestheria concinna.* The age of the formation is Neocomian.

The Zaanshiree Formation (400-600 m) conformably overlies the Khongorkhairkhan Formation, and is characterized by basalt, andesite-basalt, sometimes by the alternation of acidic effusive rocks and sandstone. The upper part is predominated by tuff, sandstone and aleurite layers containing turtles *Hangaiemys,* which indicate the late Neocomian age.

The Bayankhaan Formation (200-400 m) is composed of gray-green sandstone, siltstone, claystone and sometimes conglomerate layers. On the basis of lithological character and stratigraphical position, this formation is referred to Aptian-Albian.

2.2.3. Choir-Nyalga basin

The Cretaceous sequence of the Choir Hyalga basin is characterized by the Dorgot, Kherlen, Nyalga, Choir and Uvdugkhudag Formations, which are locally distributed in the Kherlen River basin, Gun Tsagaan nuur area, Uvdug khudag and Chandgan talyn coal mines as well as in the Khumuultiin ovoo, Jargalant, Tsaidam, Tugrug, and Tsats tsagaan Lakes areas [6].

The Dorgot Formation (80-250 m) conformably overlies the Upper Jurassic sediments, and consists of bright gray, white-green carbonaceous slate, sandstone, siltstone, clay shale, and marl layers. Remains of the early Neocomian insects Chaoberidae and Ichneumonidae; conchostracans *Esthenina* sp., *Nestoria orbicularis, Loxomicroglypta butiaetruss, Lioestheria* cf. *Imperfecta* and *Pseodoestheria concinna* were found.

The Kherlen Formation (200-1000 m) is composed of andesite-basalt, basalt and their tuffs, siltstone, claystone and sandstone. From this formation the conchostracans *Bairdestheria mediales, B. huzitai, Polygrapta* sp., *Pseudestheria* sp., and ostracods *Timiriasevia polymorpha* were identified. On the basis of these fossil occurrences, the age is considered to be Valanginian.

The Nyalga Formation (50-60 m) lies conformably on the Kherlen Formation, and is characterized by bitumen-bearing clay, paper shale, siltstone, claystone, limestone, and dolomite layers that contain the Hauterivian-Barremian molluscs *Campeloma fani, C. conica, C. asiatica, Viviparus* sp., *Probaicalia hydrobioides, Limnocyrena wagshihensisn, L. vomista, Valvata suturalis, Daurina* cf. *marginata*; ostracods *Limnocypridea abscondida, L. tumulosa, Cypridea flekodorsalis, C. unicostata, C. mundula, C. trapezoides,* and *Mongolianella palmosa*; phyllopods *Brachygrapta sibirica* and *Bairdestheria* sp.; fish *Lycoptera fragilis.*

The Choir Formation (100-200 m) unconformably overlies the Nyalga Formation, and consists of white gray, yellow gravelite, sandstone, black-gray sometimes coaly siltstone, clay, and claystone yielding mollusca *Limnocyrena sibirica, L. parva, Campeloma yihsiensis* and *Unio* cf. *huganensis*; ostracods *Cypridae concina, C.* ex gr. *prognata, C. vitimensis* and *Lycopterocypris miltifera;* charophita *Aclistochara* aff. *lata*, and dinosaurs *Psittacosaurus* sp. On the basis of fossil remains the age was referred to Aptian-Albian.

The Uvdugkhudag Formation (70-400 m) rests on the Choir Formation conformably, and is made up of black-gray conglomerate, gravelite, sandstone, coaly aleurite, and slate with intercalated coal beds that contain the molluscs *Limnocyrena altiformis, L. shantugensis,* and *L.* cf. *subplana*; ostracods *Cypridea unicostata, C.* ex gr. *concina, C. osodoevi, Timiriasevia principalis, Lycopterocypris multifera,* and *Rhiocypris* ex gr. *jurassisa*; charophita *Aclistochara* cf. *caii*; palynomorphs *Disaccites* sp. (70-80%), *Podocarpidites* sp. (1-5%), *Ginkgocycadiphytus* sp. (0-2%), *Araucariacites* (0-6%), *Tsugaepollenites mesosoicus* (0-1%), *Cyathidites minor* (2-30%), *Osmundacidites* sp. (21%), and *Polypodites minor* (18%). The age is Aptian-Albian.

The Cretaceous sediments distributed in the northern part of the Jargalant Lake area are the Tsagaantsav and Shinekhudag Formations [7]. The Tsagaantsav Formation unconformably overlies the Upper Jurassic Sharilin Formation, and consists of well-sorted and fine-grained sandstone with tuff beds. The thickness is more than 17 m.

The Shinekhudag Formation conformably overlies the Tsagaantsav Formation, and is composed of sandstone with coal seams in the lower member, monotonous mudstone in the middle member, and white sandstone with thin mudstone layers in the upper member. The total thickness is more than 80 m. The Shinekhudag Formation contains fish *Ganoides* sp. A, turtles *Asiachelus* sp., *Batoremus* sp., and *Changaiemus hoburensis*; champsosaurs *Tshoiroina nomsrai*; dinosaur *Harpimimus okladnikovi, Iguanodon orientalis,* and *Psittacosaurus mongoliensis*; crocodile Crocodylid gen. et sp. indet.; ostracods *Darwinula ovata, Cypridea zagustaica, C. spiningera, C. kosculensis, C. grandicula,* and *Lycopterocypris* ex gr. *Infantilis.*

The Shinekhudag Formation is correlated with the Hauterivian to Barremian, and the underlying Tsagaantsav Formation is judged to be approximately Valanginian to Hauterivian in age. Previous authors [8, 9] reported the age of the fossil-bearing sediments as Aptian-Albian based on pollen and molluscan assemblages, and referred it as the Khukhteeg Formation.

2.2.4. Choibalsan basin

The Choibalsan basin includes the Cretaceous Ochirkhureet, Bayantumen, Sumiinnuur, Gurvan zagal and Aduunchuluu Formations, which are exposed in the Gurvan zagal, Sumiin nuur and Choibalsan depressions [6].

The Ochirkhureet Formation (50-450 m) rests on the Upper Jurassic conformably and underlies the Bayantumen Formation. It is composed of white clay, claystone, siliceous limestone and the conglomerate layers in some places. On the basis of mollusca *Limnocyrena wangshihensis*; conchostracans *Pseudoestheria consinna, Nestoria orbicularis,* and *Bairdestheria subelongata*; insects *Chirinomaptera gobiensis, Copteclava longipoda,* and *Felundustia minaxs,* the age of enclosing rocks is assigned to the early Neocomian.

The Bayantumen Formation (500-2000 m) is distributed in the Sumiin nuur area and covers conformably the Ochirkhureet Formation. It is characterized by rhyolite, trachybasalt, ignimbrite welded tuff, perlite, andesite-basalt, tuff, claystone and shale-bearing clay containing mollusca *Limnocyrena elongata, L. sublana, Campolema conica,* and *Arguniella* sp.; conchostracans *Bairdestheria huzitai, B. sinensis,* and *Brachugrapta sibirica*; fish *Lecopteria fralis*; insect *Terrindusia reisi.* The age is considered as the middle Neocomian.

The Sumiinnuur Formation (150-400 m) lies under the Gurvanzagal Formation conformably. It is composed of alternating paper shale and clay shale, dark gray sandstone, siltstone, limestone, marl and conglomerate layers that contain the late Neocomian molluscs such as *Limnocyrena sibirica, L. wangshihensis, Viviparus turgensis, V. robustus,* and *Probaicalia vitimensis*; fish such as *Lycopteria middendorfii*; conchostracans such as *Bairdestheria middendorfii*; insect such as *Ephemeropsis triseralis*; ostracods such as *Limnocypridea grammi* and *Darwinulla contracta,* and charophita *Gouldina copacta.*

The Gurvanzagal Formation (100-350 m) underlies the Aduunchuluu Formation, and is characterized by alternating dark gray siltstone, clay, sandstone, conglomerate, and gravelite layers with molluscs *Limnocyrena altiformis, L. tani*; fossil plant *Cniopteric setacea, Czekanowskia* ex gr. *rigida,* and *Podozamites* sp. The age is Barremian-Aptian.

The Aduunchuluu Formation (50-200 m) outcrops in the depression of the Aduunchuluu coal mine and consists of light, white-gray sandstone, dark siltstone, claystone, coaly shale and coal with Aptian-Albian molluscs *Unio* sp. and *Limnocyrena altiformis*; and fossil plants such as *Birisia* ex gr. *alata, Phoenicopsis angutifolia,* and *Pityophyllum nordenskioldii.*

2.3. Gobi province

2.3.1. Basins of Transaltai Gobi

In the basins of Transaltai gobi the Lower Cretaceous is divided into the Dushuul and Baruunbayan Formations, and the Upper Cretaceous is divided into the Bayanshiree, Baruungoyot and Nemegt Formations. They widely outcrop in the Ingenii khuubur, Bugiin tsav, Shireegiin gashuun and Nemegt basins.

The Dushuul Formation (950-1000 m) overlies with angular unconformity the Paleozoic rocks and is characterized by reddish-brown conglomerate, gritstone, gray sandstone, clay, and aleurite alternating with gypsum. From this formation were identified the Aptian-Albian mollusca *Musculiopsis reshetovi, Shaerium albicum, Corbicula anderssoni, Protelliptio mongolensis, Plicatella robusta, Sainshandia dushulensis, S. mongolica,* and *S. ongonulensis,* and turtle *Hangaiemys hoburensis.*

The Baruunbayan Formation rests with angular unconformity on beds of the Dushuul Formation, and is represented by ill-sorted red-brown conglomerate and sandstone with carbonate concretions. In the upper part the black basalt layers, 30 m thick, exist. The big and small size dinosaur eggs, fragmentary bones of turtles, dinosaurs, and shells of mollusca *Pseudohyria* sp. were found.

The Bayanshiree Formation (60-70 m) covers the Baruunbayan Formation. The lower part is characterized by gray sandstones and gritstones with intercalated beds of clay, and the upper part is predominated by variegated clay and sandstones with some intercalated beds of marls and limestones, which contain the ostracods *Timiriaseva polymorpha, Cypridea cavernosa, C. fracta, Candoniella altanica, Cypris elata, Lycopterocypris baishinstavica, Mongolocypris distributa, L. bagatarachensis, Clinocypris* aff. *dentiformis, Mediocypris ectypa,* and *Talicypridea longiscula.*

The Baruungoyot Formation (45-50 m) rests on the Bayanshiree Formation without visible unconformity, and mainly consists of red sandstones with intercalated beds of aleurite and clay, lenses of carbonate rocks and conglomerates, which are rich in fossils. Dinosaurs such as *Velociraptor* sp., *Oviraptor philoceratops, Ingenia yanshini, Carnosauria* indet., *Bagaceratops rozhdestvenskyi, Saichania chulsanensis, Tachia gigantea, Therizinosaurus cheloniformis, "Dyoplosaurus" giganteus*; turtles such as *Neurankylus* sp.; lizards such as *Eoxanta lacertifrons, Globaura venusta, Pleurodontagama aenigmatoides, Priscagama gobiensis, Carolina intermedia, Cherminotus longifrons, Altanteius facilis, Gobinatus arenosus, Parameiva oculea, Tchingisaurus multivagus, Mongolochamops facilis, Mongolochamops reshetovi, Piramicephalosaurus cherminicus, Darchansaurus estesi, Macrocephalosaurus gilmorei, Cherminsaurus kozlowskii, Gobiderma pulchra, Slavoia darevskii, Adamisaurus magnidentatus,* and *Erdenetesaurus robinsonae*; amphibian such as *Gobiates khermeentsavi* and *Eopelobates leptocolatus*; and ostracods such as *Limnocythere barungoiotensis, Cypria elata, Cypridea cavernosa, C. obliquecostae, C. profusa, C.* ex gr. *fracta,* and *Mongolianella cuspidigera* were identified.

The Nemegt Formation (45-50 m) rests on the Baruungoyot Formation conformably and underlies the Naranbulag Formation. It is characterized by light gray sandstone, conglomerates, and aleurites with carbonate layers, containing many fossils. They are the turtle *Trionyx* sp., *Mongolemys* sp., *M. elegans, Basilemus* sp., *Gravemys barsboldi,* and *Yumenemys inflatus*; crocodiles *Shamosuchus ancertralis, S. tersus,* and *S. ulanicus*; dinosaurs *Adasaurus mongoliensis, Veloceraptor sp., Borogovia grasilicrus, Saurornithoides junior, Elmisaurus rarus, Oviraptor mongoliensis, Anserimimus planinychus, Gallimimus bullatus, Deinocheirus mirificus, Therizinosaurus cheloniformis, Avimimus portentosus, Alioramus remotus, "Gorgosaurus" novojilovi, Tarbosaurus baatar, Emegtosaurus mongoliensis, Opishocoelicaudia skazynskii, Homalocephale calathocercos, Prenocephale prenes, Tarchia gigantea,* and *Saurolophus angustirostris*; mammalia *Buginbaatar transaltaiensis*; and ostracods *Limnocythere barungoiotensis, Timiriasevia polymorpha, Rhinocypris ingenica, Cypridea cavernosa, C.* ex gr. *fracta, C. profusa, C. hongiloensis, C. halzanensis, Mongolocypris distributa, M. mongolica, Talicypridea biformata, Mongolinella cuspidigera,* and *Ziziphocypris martinsoni.*

The Nogoontsav Formation (= Nemegt Formation by Shuvalov [9]; Upper Nemegt Formation by Khand [10]) (20 m thick) is distributed in the Nogoon Tsav and Bugiin Tsav depressions and conformably overlies the Nemegt Formation, underlying unconformably the Naranbulag Formation. It is composed of alternating gray sandstone, reddish-brown and green clay, and carbonaceous rocks. Mammalia *Buginbaatar transaltaensis,* mollusca *Mesolanites efremovi,* and ostracods *Cypridea cavernosa, Candoniella altanica, Talicypridea biformata, T. obliquocostca, Altanicypris szczechurae, A. nogontsavica, Khandia stankevitchae, Bogdocypris ongoniensis, Eucypris tostoensis,* and *E. tooroiensis* are obtained from this formation.

2.3.2. Basins of Umnogobi (Southern Gobi)

The Cretaceous strata of these basins are relatively complete in succession and variable in sedimentary facies. The Cretaceous sediments were widespread in the Ulaan nuur, Borzongiin gobi, Mandal ovoo, Khurmen, Tsogt ovoo, Shovokht and Arts Bogd depressions. The Lower Cretaceous in this area is divided into the Tsagaangol, Ulaandel Formations and the Baruunbayan Formation, which includes part of the Upper Cretaceous. The Upper Cretaceous is divided into the Bayabshiree, Baruungoyot and Nemegt Formations [3].

The Tsagaangol Formation (800 m) is distributed near the mountain Zuramtai, in the northern part of the Borzongiin Gobi area and rests on the Paleozoic deposits unconformably. It is composed of variegated sandstone, claystone and clay, which contain the ostracods *Cypridea trita, C. remota,* and *C. prima.* The fossil records indicate the Tithonian-Barremian age. The correlative strata are both the Tsagaantsav and Shinekhudag Formations.

The Ulaandel Formation (1000-1200 m) unconformably overlies the Upper Jurassic red beds and underlies the red beds of the Baruunbayan Formation. The lower part is characterized by gray conglomerates and sandstones, and the upper part consists of gray clay and sandstone with intercalated beds of coaly claystones. In the different parts Aptian-Albian mollusca *Daurinia scalaris, Sphaeridium dowlingi, Subtilia rotunda, Viviparus keisyoensis, Bulimus rakutoensis,* and *Campeloma* cf. *yihsiensis* were found.

The Baruunbayan Formation (200 m) covers unconformably the Lower Cretaceous Aptian-Albian sediments and conformably underlies the gray beds of the Bayanshiree Formation. In some places the Baruunbayan Formation occurs on Paleozoic sediments and is covered by red beds of the Baruungoyot Formation. The lower part is composed of reddish-brown and gray-brown conglomerate-breccia and sandstones, and the upper part is charcterized by dark gray and reddish-brown clay with intercalated conglomerates and sandstones, and with abundant marl concretations. The formation yields mollusca *Bellamya matumotoi, Campeloma yihsiensis* and *Limnocyrena* sp., ostracods *Cypridea rostrata, C. prognata, C.* aff. *spinigera, C. occolata, Darwinula* sp., *Timiriasevia* aff. *grata,* and *Rhinocypris panosa,* and charophita *Mesochara (?) sainshandica.* At the Uush mountain the red beds of the Baruunbayan Formation is covered by basalts. The K-Ar isotopic dating of this basalt yields an age of 126 ± 3 Ma, and we cannot put this red-colored and coarse-grained sediments into the upper part of the Lower Cretaceous. Some authors [11] established in this basin a new formation named the Ulaanuush Formation. Thus, in view of basalt dating, the volcanic formation of this area was formed in the Neocomian Stage and the isotopic age of the basalt is not consistent with the previous age assignment.

The Bayanshiree Formation (70-100 m) is composed in the lower part of gray sandstones and gravels with some clay and conglomerate intercalation. The upper part is characterized by variegated clay and sandstones.

The Djadokhta Formation (= Djadohta Formation by Berky and Morris [4]; Bayan Dzak Formation by Barsbold [12]), 60-70 m thick, consists of red, well-sorted sand with some clay and pseudogravels intercalation. Fossil criteria indicate that the formation is Campanian in age. Remains of lizards such as *Adamisaurus mafnidentatus, Macrocephalosaurus ferrugenous, Conicodontosaurus djadochtaensis, Globaura venusta, Flaviagama dzerzhiniskii, Mimeosaurus crassus, M. tugrikinensis, Bainduis parvus, Telmasaurus grangeri, Shamosuchus djadochtaensis,* and *Gobiosuchus kielanae*; dinosaurs such as *Veloceraptor mongoliensis, Saurornithoides mongoliensis, Oviraptor philoceratops, Protoceratops andrewsi,* and *Pinacosaurus grangeri*; mammals such as *Catopsalis matthewi, Kamptobaatar kuczynskii, Bulganbaatar nemegtbaataroides, Kryptobaatar dashzevegi, Tugrigbaatar saichanensis, Sloanbaatar mirabilis, Kennalestes gobiensis, Zalambdalestes lechei, Hyotheridium dobsoni, Deltatheroides cretacicus,* and *Deltatheridium pretrituberculare*; ostracods such as *Limnocythere barungoiotensis, L. ulannurensis, L. bulganensis, Gobiocypris tugrigensis, Ilyocypris protea, Cypria elata, Lycopterocypris baga-tarachensis,* and *Bogdocypris khosbajari* were determined.

No physical contact between the Djadokhta and Baruungoyot Formations has yet been discovered. Comparative studies of dinosaurs and mammals suggest that the Djadokhta Formation is older than the Baruungoyot Formation [13]. The division of Gradzinski et al.[13] has been questioned by the team of the Mongolian Academy-American Museum Project. Dashzeveg et al. [14] stated that sampling at Ukhaa Tolgod as well as ongoing work at other localities blurs the distinction between the Djadokhta assemblages and the allegedly slightly younger Baruungoyot assemblages. Loope et al. [15] followed by Lillegraven and McKenna [16] assigned the age of this formation to late Campanian-early Maastrichtian. In the present account we continue to recognize both the Djadokhta and Baruungoyot Formations with assignment pending further publications.

The Nemegt Formation (55-60 m) conformably overlies the red beds of the Baruungoyot Formation and is characterized by dominance of gray, green-gray rhythmically alternating clay and sandstones with gravels and conglomerates.

The Lower Cretaceous deposits of the Mandal ovoo depression were classified by Khosbayar [6] into the Takhiinshand, Olonovoot, Kharmagtai, Durvendert and Tsogtovoo Formations as well as the Upper Cretaceous Mandalovoo, Ulaansuvraga and Tsagaansuvraga, but no age determination and description were done.

According to Lopatin [17], Lower Cretaceous sediments in the Shovokht depression are divided into the Mogoit, Manlai, Shinekhudag, Khukhteeg, and Sainshand Formations.

One of the complete sections, in which continuous sedimentation is well recorded, is the

Undurbogd area. It was suggested as a key section for this basin, where the Tsagaantsav, Manlai, Shinekhudag, Khukhteeg, Baruunbayan, Bayanshiree and Bayanzag Formations were identified on the basis of the ostracod assemblages [18]. The absolute data from basalt overlying the red beds of the Bayanzag Formation are 71.6 ± 1.6 to 73.5 ± 1.6 Ma by K-Ar methods (pers. com., Watabe).

2.3.3. Basins of Dornogobi (Eastern Gobi)

According to Shuvalov [3], in this basin the Lower Cretaceous Tsagaantsav, Shinekhudag, Khukhteeg and Baruunbayan Formations and the Upper Cretaceous Bayanshiree, Baruungoyot and Nemegt Formations were determined. The Cretaceous strata are well developed in the Ulgii khiid, Zuunbayan, Sainshand, Shazangiin Gobi, Unegt and Khar Khutul areas.

The Tsagaantsav Formation conformably overlies the Upper Jurassic Shariliin Formation and is distributed in all depressions. The total thickness is about 250-1100 m. The lower part is characterized by conglomerate, gravelite and basalt alternating with trahcybasalt tuff, and the upper part is composed of tuff, sandstone, slate clay, siltstone, and claystone alternation containing mollusca *Mycetopus transbaicalensis* and *M. qyadratus*; ostracods *Mongolianella palmosa, Darwinula contracta, Timiriaseva polymorpha, Cypridea trita, C. prima,* conchostracans *Estheria middendorfii,* and *E. amurensis*; fish *Lycoptera middendorfii.*

Originally the Tsagaantsav Formatiom was established in the basin of Dornogobi at the Tsagaan Tsav well by Vasilev et al. [19]. Later Martinson et al. [20] changed the age to the Upper Jurassic-Lower Cretaceous (Tithonian-Valanginian). According to Shuvalov [3], age of the Tsagaantsav Formation is Valanginian-Berriasian.

The Shinekhudag Formation [20], 50 m thick, is well developed in this basin and conformably overlies the Tsagaantsav Formation, and consists of gray, gray green conglomerate, gravelite, sandstone, clay, bitumen-bearing shales, paper shales, and marls containing mollusca *Limnosyrena pusilla, L. mongolica, L. altiformis, Liolax parva,* and *Bithunia dzunbaiensis*; ostracods *Cypridea oriformis, C. osodovi, C. aragangensis, Phinocypris tugurigensis*; conchostracans *Bairdestheria argunica, B. chinchudukensis, Pseudograpta khalfini*; fish *Lycoptera fragilis* and *Stichopteus reissi*; and charophyta *Mesochara voluta.* The fossil record indicates the age as Hauterivian-Barremian. According to the Vasilev et al. [19], it was the lower Zuunbayan Formation.

The Khukhteeg Formation [20], 250-470 m thick, with angular unconformity overlies the Shinekhudag Formation, and is composed of green-gray aleurite, clay and limestone with conglomerate and sandstone intercalation. Mollusca *Viviparus obustus, Protelliptio hamili, Unio paletsensis, Limnocyrena elliptiformis,* and *L. anderssoni*; dinosaurs *Iguanodon orientalis*; phyllopods *Estherites shanchoensis* and *E. orthothemoides* occur in this formation. The

Khukhteeg Formation is considered to be Aptian-Albian in age. According to Vasilev et al. [19], it was the upper Zuunbayan Formation.

The Baruunbayan Formation is composed of red-colored conglomerates with intercalated lenses and layers of gravelites, various grained sandstone and clay debris. The thickness ranges from 50 to 200 m.

The Bayanshiree Formation (150-250 m) conformably overlies the Baruunbayan Formation and is distributed almost in all the eastern Gobi depressions. Typically it is predominated by gray, sometimes variegated sandstones and gravelites in the lower part and by clay with sandstone and marl layers in the upper part, yielding ostracods *Talicypridea longiuscula, T. biformata, Lycopterocypris bajshintsavica, L. khandae, Cypria elata, Cypridea* aff. *profusa, Timiriasevia polymorpha, T. martinsoni, Mongolianella cuspidigera, Cyclocypris transitoria,* and *Candoniella altanica*; mollusca *Pseudohyria cardiiformis, P. hongilica, P.* cf. *tuberculata, Pseudohyria* sp., and *Sainshandia* (?) sp.; turtles *Trionyx* sp. Fossil evidence suggests that the Bayanshiree Formation is Cenomanian-Santonian in age.

The Javkhlant Formation (Baruungoyot Formation by Shuvalov and Stankevich [21]), 15-60 m thick, occurs on the Bayanshiree Formation, and is represented by red sandstones and sandy clay, alternating mainly with gravelite, and sometimes with conglomerate and sandstone concretion. The formation contains ostracods *Cypridea* aff. *profusa, C. cavernosa, C. distributa, Lycopterocypris bajshintsavica, Mongolianella cuspidigera, Cypria elata, Candoniella altanica, C. bugintsavica, Cyclocypris transitoria* and *T. biformata,* and dinosaurs from the family Ornithomimidae. The fossil horizon of the Javkhlant Formation is correlative with the Santonian-Campanian Stage.

The Nemegt Formation (up to 70 m) overlies the Javkhlant Formation and is characterized by variegated, various grained, poorly cemented sandstone alternating with reddish-brown and yellow-gray sandy clay, marl concretions, gravelite and limestone. The formation yields a large and diversified fauna such as bones of different kinds of dinosaurs, crocodiles *Paralligator* sp; turtle *Mongolemys* sp.; mollusca *Mesolanistes efremovi, M. mongolensis,* and *M. shuwalovi,* and ostracods *Cypridea distributa, C. barsboldi, Candoniella altanica, Cyclocypris transitoria, Cypria elata, Timiriaseva* aff. *costata, Mediocypris* sp., and *Talicypridea biformata,* that appears to be of the Maastrichtian age.

2.3.4. Tamtsag depression

The stratigraphic sequence of the Tamtsag depression is basically same as that of the Dornogobi basin. Khosbayar [6] established in the Tamtsag basin the Tsagaantsav, Zuunbayan, Sainshand and Bayanshiree Formations but the description of them was not published. The Zuunbayan Formation is correlative with the Shinekhudag and Khukhteeg Formations in the basins of Dornogobi.

2.4. Conclusion

The formations discussed above represent all the known Cretaceous units of Mongolia in this first general review of the stratigraphy of the Mongolian Cretaceous (Table 1). Evaluation of formational ages is based on the 'stages of evolution' of vertebrate and invertebrate assemblages. We believe that this scheme for the Cretaceous deposits of Mongolia will provide a useful framework for future stratigraphic and paleontological work.

Upper Jurassic-Lower Cretaceous deposits of the Tsagaantsav Formation conformably overlies coarse-grained Upper Jurassic red beds in all the basins exept the Shavokht depression of Gobi province, where they overlie the Paleozoic with angular unconformity. Recently Sinitsa [5] suggested that the boundary between the Jurassic and Cretaceous Systems in the Valley of Lakes be defined by subdivisions of ostracod biostratigraphical units. On the basis of careful correlation with the international type section in Europe the boundary is placed between the *Cypridea unicostata–Torinina mongolica* and *Cypridea fasciculata–Cypridea setina* assemblage zones, corresponding to the base of the Undurukhin Formation overlying the Tevsh Member of the Dundargalant Series.

The lower part of the Upper Cretaceous was defined in the Zuunbayan depression of Gobi province as the Sainshand Formation of the Cenomanian age [22]. Later Barsbold [12], on the basis of mollusca assemblages, changed the age of the Sainshand Formation to the Albian-Cenomanian. The K-Ar dating [23] of basalt at the Khar Khutul section of the Zuun bayan depression is 145 to 119 Ma. Thus, geochronologically the above mentioned formation is correlated with the Tsagaantsav Formation. The rock unit now known as the Baruunbayan Formation was first described by Shuvalov [3] in the area of the Baruunbayan mountains in the Mandal ovoo-Ulaan nuur depression and was referred to the upper part of the Lower Cretaceous. So the lowermost part of the Upper Cretaceous became the Bayanshiree Formation. Later, Khand [10] studied ostracod assemblages from the same section and considered that the Baruunbayan Formation is transitional between the Lower and the Upper Cretaceous, and the Age is Albian-Cenomanian. The K-Ar isotopic dating of basalts overlying the Baruunbayan Formation at the Uush mountain [11] in this depression is 126 ± 3 Ma. Thus, volcanic activity in this area was during the Neocomian.

The uppermost Maastrichtian strata have not been documented in Mongolia. The Cretaceous/Paleocene contact as shown by the Naran gol section at Naran bulag of the Nemegt graben is marked by well-developed caliche paleosols overlain by an extensive erosional unconformity. Both the Cretaceous and Paleocene strata contain semiarid palaeosols. Paleonto logical evidence is not sufficient to assess the magnitude of the time gap between the Upper

Table 2
Stratigraphic position of representative Cretaceous fossil assemblages

System	Stage	Formation	Charophytes	Spores & Pollen	Plants			Molluscs	Ostracods	Dinosaurs	Conchostracan	Turtles	Lizards	Mammals
					Krassilov [29]	Sodov [30]	Makulbekov & Kurzanov [32]							
Cretaceous	Maastricht.	Nemegt			*Nyssoidea mongolica*			*Buginella buginica* *Mesolanistes*	*Talicypridea reticulata*; *Mongolocypris distributa*	*Tarbosaurus baatar* *Gallimimus bullatus* *Deinocheirus mirificus*	*Estherites* *Paleoleptestheria* *Lioestheria*	*Mongolochelus* Lindholmemydidae Haichemydidae	*Slavoia darevskii* *Shinsauroides latipalatum*	*Buginbaatar transaltaiensis*
	Campanian	Baruungoyot	*Mongololiachara* *Mesochara*				*Ginkgo* *Platanus* *Trochodendroides* *Nelumbo* *Quereuxia*	*Pseudohyria barungoiotica* *Buginella laevivalvata*	*Talicypridea abdarantica*	*Velociraptor mongoliensis* *Protoceratops andrewsi* *Monony chusolecranus* *Pinacosaurus grangeri*	?		*Priscagama gobiensis* *Mongolocha mopsfacilis*	*Nemegtbaatar gobiensis* *Deltatheridium tardum* *Asiatherium reshetovi*
	Santonian	Djadokhta			*Bothrocaryum gobience* *Nyssoidea mongolica*			*Plicatotrigonioides gobiensis*	*Gobiocypris tugrigensis*			*Adochida* Nanhsiungchelyidae	*Mimeosaurus crassus* *Gobekkocretacicus* *Isodontosaurus gracilis*	*Kamptobaatar kuczynskii* *Zalambdales teslechei* *Deltatheroides cretacicus* *Kennalestes gobiensis*
	Coniacian		?				?							
	Turonian	Bayanshiree	*Atopochara multivolvis* *Caucasuella gulistanica*					*Pseudohyria cardiiformis*	*Lycopterocypris baishintsavica*	*Segnosaurus galbinensis* *Garudimimus brevipes*	*Sphaerograpta* sp. *Eulimnadia drjachlovae*			?
	Cenoman.		*Mesochara sainshandinica* *M. tuzsoni*		?						?			
	Albian	Baruunbayan		*Foraminisporites assemmetricus* *Alisporites elongatus* *Abiespollenites* sp.	*Limnothetis* *Limnoniobe*	*Cyparacites anlajensis* *Potamogeton* sp.		*Campeloma yihsiensis* *Oxynaia sainshandica*	*Cypridea rostrata*	?	*Paleoleptestheria* sp.	?	?	*Guchinodon khoburensis* *Eobaatarmagnus* *Gobiodon infinitus* *Arguimus khosbajari* *Kielantherium gobiensis* *Plakennalestes trofimovi*
	Aptian	Khukhteeg	*Atopochara trivolvis* *Mesochara voluta* *M. tuzsoni*		*Baierella hastata* *Araucaria mongolica*	*Araucaria* sp. *Paratoxodium jacutensis* *Ginkgo adiantoides* *Asplenium disconianum*; *Muscitites ostracodiferas* *Dicksonia concinna* *Otozamites lacustris*		*Sainshandia ushlegica* *S. dushiulensis*	*Tsetsenia mira* *Trapeoidella khandae* *Cypridea acutituberculata* *Janinella tsaganensis*	*Harpymimus okladnikovi* *Iguanodon orientalis*	*Bairdestheria multa* *Estherites liushinensis*	*Hangaiemys leptis* *Asiachelys perforata*; *Hangaiemys hoburensis* Macrobaenidae *Moingolemys* sp.		
	Barrem.	Shinekhudag	*Aclistochara caii* *Raskyella* sp.	*Cicatricosisporites australiensis* *Densoisporites velatus* *Pilosisporites trichopapilosus*	*Otozamites* *Pseudolarix erensis*	*Birisia onychioides* *Ruffordia geoppertii* *Otozamites* sp. *Arzucarites* sp.			*Cypridea fasciculata*	?	*Bairdestheria memorabilis* *B. shinchudukensis*			
	Hauteriv.													
	Valang.	Tsagaantsav			*Baiera manchurica*	*Cladophiledis* *Onychiopsis* *Nilssoniopteris denticulata*			*Cypridea unicostata*	*Psittacosauridae*	*Opsipolygrapta anferovi* *Bairdestheria enucleata*			
	Berrias.													

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Cretaceous and Paleocene in the Gobi basins.

3. BIOLOGICAL ENVIRONMENTS

The nonmarine Cretaceous of Mongolia is abundant in fossil remains represented by charophytes, plants, sporopollen assemblages, fresh-water bivalves, ostracods, conchostracans, insects, fishes, amphibians, reptilians and mammals. These faunas and floras serve as one of the most important bases for stratigraphical classification and correlation. In Table 2 the stratigraphic positions of representative Cretaceous fossil assemblages are shown.

3.1. Charophytes

The Mesozoic lacustrine deposits of Mongolia are rich in fossil charophyta [24]. In Hauterivian-Barremian deposits gyrogonites of charophyta were found in the vicinity of Ugiin nuur (Shinekhudag Formation). Among them *Aclistochara caii* and *Raskyella* sp. were determined.

The Aptian-Albian assemblage of charophyta is poor in species. Only three species were determined from Mongolian Altai, Northeastern Mongolia and Southern Gobi: *Atopochara trivolvis, Mesochara voluta,* and *Mesochara tuzsoni.*

Albian-Cenomanian deposits of Southern Gobi (Sainshand Formation) contain *Mesochara sainshandinica* and *Mesochara tuzsoni.*

The assemblage of the Turonian stage from the Bayanshiree Formation in Southern Gobi also includes *Atopochara multivolvis* and *Caucasuella gulistanica.*

The Santonian-Maastrichtian Stages of the charophyta evolution in Mongolia are characterized by an abundance of *Mongolichara* (endemic genus) and *Mesochara* (Bayanzag, Baruungoyot and Nemegt Formations).

3.2. Palynomorphs

Palynological analyses were done for the Cretaceous sediments of Shivee ovoo [25], Aduunchuluu and Bayan-Erkhet [26, 27] coal-bearing localities. Aduunchuluu and Shivee ovoo localities have yielded 47 species of 46 genera, and the age of the assemblage is assumed to be Aptian and Albian. The spore-pollen assemblage of Bayan-Erkhet includes 29 species and 35 genera. Based on the composition it may be referred to the Hauterivian-Barremian in age and belongs to the Bayanerkhet Formation.

One of the more recent palynological analyses was undertaken by the Mongolia-Japan Joint Paleontological Expedition [28]. The goal was to establish age relations between sites, from which dinosaurs were being collected and to provide paleoecological data for environmental reconstructions. Units sampled include Upper Cretaceous formations exposed in several badland areas that are yielding dinosaurs and other vertebrate fossils. Oxidation and coarse grain size of Upper Cretaceous units cause them to have low potential for preservation of organic microfossils. Palynomorphs have been only recovered from outcrops of the Lower Cretaceous. These assemblages are indicative of the Aptian-Albian age.

3.3. Plants

The Lower Cretaceous lacustrine beds of Mongolia are rich in faunal remains, but the fossil plants are scarce. Rich plant localities were discovered by the Soviet-Mongolian Geological Expedition in the Central Mongolia region, where plant-bearing beds occur in the Undurukhaa, Andakhudag and Khulsangol Formations, while the correlatives in the western and eastern areas are assigned to the Gurvanereen, Tsagaantsav and Shinekhudag Formations [29]. Four assemblage zones are recognized. Two lower zones are assigned to the Neocomian, and next two zones to the Aptian. These age assignments essentially agree with the invertebrate data.

Plant megafossils came from the tuffaceous beds and aleuropelitic paper shales–varvites, deposited in meromictic lakes. Varvites contain abundant pollen grains in the lighter colored lamellae. There are also coprolites and caddis fly cases with plant remains. Aquatic Lycopsida are abundant in the upper horizons, possibly evidencing the eutrophication of the lakes.

At the beginning of the Cretaceous the fern marshes were drastically reduced. Mud flats around the lakes were colonized by the hydrophylic bennettites and presumably also by the primaeval reed- and sedge-like monocots. Other angiosperms represented by winged fruits might grow on slopes occupied by the *Pseudolarix* forests. The *Araucaria–Brachyphyllum* lowland community is conceived as the eastern extension of the European–Central Asian brachyphyllous forests, while some Siberian refugees, supposedly confined to the slopes, have given the Mongolian flora an ecotonal aspect.

Recently the following five palynological assemblages were determined in the Lower Cretaceous of Mongolia [30]: Berriasian–early Hauterivian, late Hauterivian–Barremian, late Barremian–Aptian, late Aptian–early Albian and Albian. Differences from the Krassilov's assemblages [29] are shown in Table 2.

In the locality Khar Khutul, sediments of the Bayanshiree Formation (Turonian–Santonian) contain *Bothrocaryum gobience* and *Nyssoidea mongolica* [31]. The latter was also found from the Nemegt Formation (Campanian-Maastrichtian) at the locality Bugiin Tsav. Recent North American species of Nyssa grows mainly along the river bank in humid warm climate.

The Cenomanian–Turonian flora [32] from Mongolia has not been confirmed yet. The Senonian flora is poorly known. The Santonian–Campanian deposits of the Baruungoyot Formation in the Altan uul-4 yield the prints of leaves of *Ginkgo, Platanus, Trochodendroides, Nelumbo* and *Quereuxia.*

3.4. Molluscs

Cretaceous fresh-water molluscs were studied in detail by Barsbold [12] and Martinson [33]. Recently Sersmaa [34] summarized their data. Fresh-water molluscs were found from 44 localities in southern Mongolia. 84 species belonging to 34 genera were discovered, among which 26 species are from the Lower Cretaceous, 54 species from the Upper Cretaceous, and 4 species from the Lower–Upper Cretaceous transitional deposits.

In Cretaceous sediments five assemblages of fresh-water molluscs were identified by Martinson [33]: Dushuulian (Albian), Sainshandian (late Albian–early Cenomanian), Bayanshireenian (late Cenomanian–Santonian), Baruungoyotian (Campanian) and Nemegtian (Maastrichtian).

3.5. Ostracods

The stages established by nonmarine ostracods facilitate recognition of local stratigraphical divisions in Mongolia [10]. Stratigraphical studies of the Cretaceous nonmarine ostracods of Mongolia have revealed significant temporal changes and geographical heterogeneity. The Lower Cretaceous stage represented by the Tsagaantsav, Shinekhudag and Khukhteeg Formations and the Upper Cretaceous stage represented by the Baruunbayan (*Cypridea rostrata*), Bayanshiree (*Lycopterocypris baishintsavica*), Bayanzag (*Gobiocypris tugrigensis*), Baruungoyot (*Talicypridea abdarantica*), lower Nemegt (*Mongolocypris distributa)* and upper Nemegt (*Talicypridea reticulata*) Formations. The best environmental conditions for ostracods existed during the deposition of the Shinekhudag and Nemegt Formations.

Since the beginning of the Early Cretaceous, intensive development of nonmarine ostracods took place. During the Early Cretaceous, such genera as *Timiriasevia, Theriosynoecum, Limnocypridea, Cypridea, Mongolianella, Hyocyprimorpha* and *Lycopterocypris* were widespread, of which by the end of the Early Cretaceous *Theriosynoecum, Limnocypridea* and *Ilyocyprimorpha* had disappeared. There was also a marked drop in species diversity of the genera *Timiriasevia, Theriosynoecum, Limnocypridea, Cypridae, Mongolianella, Lycopterocypris* and some others, but some species belonging to these genera survived into the Late Cretaceous.

During the early Late Cretaceous, the ostracod faunas were still much common with Early Cretaceous ones, both on the genera and, to some extent, on the species level. However, at the beginning of the Bayanshiree Formation new genera appeared, and some characteristic genera of the Late Cretaceous faunas (e.g. *Mongolocypris, Talicypridea, Gobiocypris, Altanicypris*) have survived into modern faunas (*Limnocythere, Ilyocypris, Eucypris, Candona, Cypria, Cyclocypris, Cyprinotus*). During the Late Cretaceous, extinctions and new appearance of ostracod genera and species continued, although generally the Upper Cretaceous is characterized by relatively gradual development of the ostracod faunas, reflecting more or less stable conditions in the lake basins of the Gobi during that interval.

According to the study of the biostratigraphical and lithostratigraphical units in Central Mongolia [5], the Undurukhaa Formation and Boontsagaan Series were assigned to the Lower Cretaceous. The ostracod assemblages *Cypridea unicostata–Torinina mongolica* characterized the Tsagaantsav Formation (analogue of the Andakhudag, Builyastyn and Shandgol Beds) and belongs to the Upper Jurassic. The assemblages *Cypridea fasciculata–Cypridea setina* represented the Lower Cretaceous in the ostracod development and referred to the Shinekhudag Formation. Two other Early Cretaceous assemblages, namely, *Cypridea acutituberculata–Janinella tsaganensis* and *Tsetsenia mira–Trapezoidella khandae,* distributed only in this region and have no analogue at other basins of Mongolia.

3.6. Conchostracans

Conchostracans were well adapted to the periodically drained environments and well documented in the Lower Cretaceous deposits of Mongolia. Early Cretaceous conchostracans of Western, Central and Eastern Mongolia were studied by Shuvalov and Trussova [35], and six conchostracan complexes were established, successively changing each other and corresponding to the lower Tsagaantsavian (Upper Jurassic-lowest Valanginian), upper Tsagaantsavian (Valanginian), lower Shinekhudagian (Hauterivian), upper Shinekhudagian (Barremian), lower Khukhteegian (Aptian-Albian), and upper Khukhteegian (? Albian) biostratigraphical units.

3.7. Insects

The Lower Cretaceous beds of Mongolia, especially the paper shale facies are famous for their exceptionally rich fauna of crustaceans, insects and fishes. These beds have been extensively studied by the Joint Soviet-Mongolian Paleontological Expedition in the area ranging from the Great Lakes in the west to the eastern Gobi. In three books "Early Cretaceous lake Manlai" (edited by Tatarinov, 1980), "Insects in the Early Cretaceous ecosystems of Western Mongolia" (edited by Tatarinov, 1986) and "Jurassic and Lower Cretaceous of Central Mongolia" [5], most of known insects have been published, but no stratigraphical subdivisions have been suggested.

3.8. Turtles

According to Narmandakh [36], two turtle assemblages are recognized in the Lower Cretaceous. The lower one characterized the lower part of the Dushuul Formation, and is characterized by *Hangaiemys hoburensis*, by the early taxa of Macrobaenidae, and undescribed *Mongolemys* sp. The upper one is found in the upper part of the Dushuul Formation (analogue of Khukhteeg) and represented by *Hangaiemys leptis*, by a new species *Asiachelys perforata* from one family. The species *Pteishanemys testudiformis* is only one representative from the family Sinochelyidae in Mongolia, which is close to the Early Cretaceous turtle from China.

The following four assemblages: lower Bayanshiree, upper Bayanshiree, lower Baruungoyot and upper Baruungoyot definitely differ from the lower and upper assemblages. The turtle assemblages from there are represented mainly by large *Adochida* and taxa from the *Basilemys* group–Nanhsiungchelyidae.

The turtle assemblage from the Nemegt Formation shows greater diversification. The change in the depositional setting coursed to the mass death, sometimes forming the rock, for example at Bambu khudag, Tsagaan khushuu and Bugiin tsav. The large *Adochida* and Nanhsiungchelyidae completly disappeared. Turtle from the group *Mongolochelus* with water adaptation and close to the Early Jurassic form from North America became the dominant fossil. Lindholmemydidae was replaced by turtles from the genus *Mongolemys*. Also a new endemic family Haichemydidae appeared, which was very close to *Mongolemys*, but with more progressive features of skull and carapace and plastrone join.

3.9. Lizards

The list of fossil lizard taxa from the Cretaceous of Mongolia was given by Jerzykiewicz and Russell [37] and then revised by Borsuk-Bialynicka [38]. She named more correctly the lizards from the following formations: Djadokhta, Baruungoyot and Nemegt. The latter is referred to the Nemegtian age and the former two to the Baruungoyotian age.

Alifanov [39] on the basis of study of Teioidea established four stages in the lizard evolution during the Cretaceous. According to him, the stratigraphical distribution of the Cretaceous lizard families are: Upper Jurassic–Lower Cretaceous, Hauterivian–Barremian included 9 families, upper Santonian–middle Campanian consisted of 14 families, and lower Maastrichtian composed of 4 families.

3.10. Dinosaurs

Cretaceous strata in Mongolia contain abundant fossil faunas, including numerous world famous dinosaurs. Since the first discovery, Mongolia has contributed more than 50 genera to the world`s dinosaur treasury, making up almost 20 % of all valid genera of dinosaurs currently known. Most of these taxa are based on relatively complete specimens, and some are represented by hundreds of individuals. Mongolian dinosaurs are known from a comparatively narrow time interval, mainly from the Late Cretaceous, although some have been found in the Early Cretaceous [40].

The oldest Cretaceous dinosaur fossils were found from the Tsagaantsav Formation (supposedly Berriasian–Valanginian) and are represented by fragments of Psittacosauridae.

The first known carnivorous dinosaurs comes from the Shinekhudag Formation (Hauterivian–Barremian). A few species of dinosaurs have been found from this formation: *Harpymimus, Psittacosaurus,* and *"Iguanodon"*.

Dinosaurs from the early Late Cretaceous (Baruungayan Formation) are poorly known everywhere. The environment was becoming progressively drier during this time.

Among the Bayanshiree Formation (Cenomanian-Turonian) discoveries the *Segnosaurus* and *Erlicosaurus* [41, 42] remains are interesting. Also, Therizinosaurs, Ornithomimids and indefinite sauropods were found.

Central Asia became more arid during the Late Cretaceous time. The Djadokhta and Baruungoyot Formations are of this age: the Djadokhta Formation (lower Senonian) is characterized by Dromaeozaurids, Saurornithoides, Oviraptorids, Pinacosaurus, Hadrosaur babies, and Protoceratopsids; the Baruungoyot Formation (upper Senonian) by Oviraptorids, Ankylosaurs, Pachycephalosaurs and Bagaceratops.

Central Asia experienced more moist conditions about 70 million years ago. In the Nemegt depression of southwestern Mongolia interconnected river channels developed in the inter-dune valleys. Nemegtian Oviraptorids, Therizinosaurus, Deinocheirus, Mononykus, Thyrannosaurids, Hadrosaurs, Saichania and other Ankylosaurs, and Pachycephalosaurs are known. Dinosaur nests were first discovered in Mongolia in 1922 by the Third Central Asiatic Expedition of the American Museum of Natural History. Nesting sites with nests or eggs containing identifiable embryonic remains are rare. Fossil eggshell fragments are relatively common in most of the Upper Cretaceous and in two Lower Cretaceous localities. The Mongolian Academy-American Museum Project team made interesting discoveries, such as the theropod embryo [43] and a nesting oviraptor [44]. Recently the results of study of egg- shells collected during the field seasons of 1993-1996 by the Mongolia-Japan Joint Paleontological Expedition were published [45, 46].

Abundant dinosaur footprints were discovered by the Mongolia-Japan Joint Paleontological Expedition from Upper Cretaceous sandstone beds in Shar Tsav, 250 km east of Dalanzadgad, center of South Gobi. This is the second discovery of dinosaur footprints after the ambiguous short report of Cretaceous dinosaur footprints from central Mongolia. The Shar Tsav footprints allow various analyses of dinosaur kinematics and behaviour for the first time in Mongolia because of their remarkable quality of preservation and abundant quantity [47]. Some new areas with footprints were established by this expedition in the locality Avdarant, Ulaan nuur basin (pers. comm., Tsogtbaatar). Also the tracks of dinosaurs were discovered in large-scale cross-stratified eolian rocks, and their skeletons are found in dune margin alluvial fan sediments. Fans were active during mesic intervals when dunes were stabilized, tracks were preserved under xeric conditions in the deposits of rapidly migrating dunes [15].

3.11. Mammals

In the past few years Russian and American scientists working in the Gobi have uncovered preserved remains of over 200 individual mammals. The Gobi mammals, which date to about 15 million years before the extinction of the dinosaurs, are important for exploring the state of mammal diversity at the geological moment before their rapid evolutionary expansion. One important implication of the recent Gobi discoveries concerns the number of different kinds of mammals that had evolved by the Cretaceous.

The first known level is Aptian–Albian [48]. They are known from the locality Khuubur in the Valley of Lakes and represented by Early Cretaceous multituberculates (subfamily Eobaatarinae), some eutherian taxa, two taxa of the "triconodonts" (*Gobiconodon borissiaki* and *Guchinodon khoburensis*), one symmetrodonta (*Gobiodon infinitus*), two "eupantotheres" (*Arguimus khosbajari* and *Arguitherium cromptoni*) and one aegialodontids (*Kielantherium gobiensis*).

The multituberculate assemblages from the Djadokhta and Baruungoyot Formations and equivalent of the Nemegt Formation represent nine described and several undescribed species, among which the three taeniolabidoid families: Eucosmodontidae, Sloanbaataridae and Taeniolabididae are included. Late Cretaceous eutherian (four monotypic genera), as well as deltatheroidans are known from the Djadokhta and Baruungoyot Formations. The only marsupial known from the Baruungoyot Formation is *Asiatherium reshetovi*.

The mammals housed in Paleontological Institute, Russian Academy of Sciences, still need to be described. It is hoped that newly discovered material as well as ongoing anatomical studies will improve our knowledge.

4. CLIMATOLOGICAL ENVIRONMENTS

Many studies have documented that climate has a profound influence on facies distribution, sedimentation patterns, floral and faunal endemism, and provenance signatures in nonmarine sedimentary basins. However, the geological history of the basins in Mongolia has been interpreted in several ways.

The Lower Cretaceous is represented largely by lacustrine sediments. Regional syntheses of Mesozoic nonmarine strata from adjacent northern China suggest a widespread west to east increase in humidity during the Mesozoic. Although not as well documented, a similar west to east climatological trend has been reported from Mongolia. Devjatkin has documented in the Lower Cretaceous the dominance of red beds sequences in the Gurvanereen and Zereg Formations in Western Mongolia, suggestive of an arid climate. Conversely, Shuvalov [23] has described contrasting time equivalent sequences of mainly gray-green fuvial-lacustrine oil-shale and coal-bearing sequences, suggestive of a humid climate, in Eastern Mongolia. According to Hendrix et al. [49], model predictions of Early Cretaceous atmospheric circulation patterns for much of Central Asia suggests northerly trade winds, similar to modern circulation patterns. Western Mongolia was the site of well-circulated regionally extensive Early Cretaceous lakes that did not preserve sufficient organic matter for source rock viability. Eastern Mongolian basins (Nyalga, Tamtsag) were the sites of regional, poorly-circulating lakes that preserve considerable quantities of organic matter and sourced the petroleum of those basins. These Cretaceous anoxic lacustrine deposits are of the combined results of regional rifting, rapid subsidence and increased accommodation space, the Early Cretaceous global warming trend, and the region's proximity to Pacific maritime moisture. Matsukawa et al. [7] described in the Choir basin depositional environments of the Shinekhudag Formation (?) represented by a fluvial system comprising meandering rivers.

Many controversial opinions have been published regarding the depositional enviroments of the Upper Cretaceous. Notwithstanding strong evidence for dune and interdune sedimentation, a lacustrine interpretation is deeply entrenched in the literature [23, 31, 32, 50, 51, 52]. Sochava [53] regards that alluvial cycles are reflected in the Bayanshiree, Baruungoyot and Nemegt Formations. Wind-blown and associated facies of subaerial deposition, interfingering with water-laid interdune/ephemeral facies, have been documented from the Upper Cretaceous Djadokhta and Baruungoyot Formations [4, 54, 55, 56]. Some of eolian facies at the Tugrigiin shiree locality have been previously interpreted as a "dry delta" in a "shallow-water, wave-dominated environment" by Tverdochlebov and Tsybin [57].

According to Jerzykiewicz [58], the sedimentological record indicates that the climate of the Gobi basins in the Late Cretaceous was semi-arid. The indications of an oscillating ground-water table in the Upper Cretaceous sections of the Gobi basins, similar to those documented in modern semiarid areas existed. The calcrete horizons in the Gobi are analogous to those described from the Kalahari beds of Namibia and Botswana, which form the bedrock of the modern Okavango Delta. In that area the calcretes are being formed in a multistage process involving alternation of climatic conditions. Setting of calcic dust into the host sediment occurs during a subterraneous stage in semiarid to subhumid conditions. Diagenesis, which is often followed by exhumation, weathering and redeposition, takes place during a subaerial stage in semiarid to arid condition.

He also mentioned conspicuous similarities between the Late Cretaceous Gobi and the present day Okavango vertebrate habitats. The most important of these similarities is an alternation of wind-blown dunes and water-laid lacustrine and fluvial deposits. Changes from arid conditions to subhumid conditions may have been gradual or instantaneous.

More recently, Loope et al. [15] have published new results of sedimentological observations on the Djadokhta Formation at Ukhaa tolgod of the Nemegt basin. They interpret the highly fossiliferous rocks at Ukhaa tolgod as non-eolian. All of the Ukhaa tolgod fossils are found in structureless sandstones, which contain pebbles and abundant coarse sand. Alluvial fan interpretation that buried dune slopes during mesic intervals provides an alternative to the "sand-storm" (eolian) hypothesis for in situ burial of articulated skeletons. The high, sparsely vegetated landforms were a sediment reservoir available for episodic gravity flows, capable of quickly burying intact, unscavenged skeletons, and, possibly, live animals.

On the basis of biogeochemical studies of palaeohydrochemistry of Cretaceous continental water, Kolesnikov [59] established that at the beginning of the Early Cretaceous the limnic basins existed in humid subtropical climate condition with average annual temperature of 17° to 20°C. The mineralization of water was not high. By the end of the Early Cretaceous the aridization of subtropical climate was observed and the seasonal contrast was increasing. At the end of Early Cretaceous (Ongon ulaan uul) the aridity persisted and mineralization of large water reservoir increased up to 6.1–10.4 %. During the Late Cretaceous the climate was arid and the maximum rise in temperature was at the Baruungoyotian time, and the salinity of basins reached 11.5–12.2 %. At the end of the Late Cretaceous at the Nemegtian time (Maastrichtian) the tropical arid climate continued and degree of mineralization went down and reached 9.8–11.3 %.

Thus, during the Cretaceous time a marked paleogeographic and climatic shift to drier conditions are observed. Oxic lacustrine and associated facies of subaqueous origin, mainly fluvial predominates in the Lower Cretaceous, changed to eolian and associated interdune facies and meandering rivers and lakes on an alluvial plain in the Upper Cretaceous.

5. CONCLUSION

In spite of relatively detailed studies on the nonmarine Cretaceous of Mongolia, the complexity of the relationship of the multifacial deposits, different age estimations established by the diverse nonmarine fossils as well as the difficulties and inaccuracies of the remote correlations made on their basis, continue to influence the making of more or less unified stratigraphical schemes. Absolute radiometric dates would significantly improve the basement of divisions and their correlations, but it is too rare and insufficient. The position of some stratigraphic divisions, for example, Bayanshiree, Baruungoyot and others, more lower divisions remain to be quite unclear. Difficulties connected with the remote correlations influence dividing of more detailed units within the Cretaceous often putting them on the level of usual and temporal conventions and suppositions.

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Paleoenvironmental changes during the Cretaceous in eastern China

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From the latest Jurassic to Early Cretaceous, volcanism was widespread in North and Southeast China extending into southern Korea and southwestern Japan. A large-scale horizontal displacement along the Tan-Lu fault zone started from the Aptian and ended within the Albian Stage with a horizontal strike-slip distance of 740 km. A seaway and possible river connected Northeast China with Japan. Due to the Yanshen Movement caused by the collision of the Himalayan and Lhasa Terranes and the westward subduction of the Pacific Plate, which greatly altered paleogeographic patterns throughout China, there were two newly formed huge lakes in the central China and Manchuria regions during the Late Cretaceous (Yunmeng and Songhua Lakes). In the meantime, the east part of the Zhejiang-Fujian-Guangdong Provinces rapidly elevated to form the Coastal Mountains ranging between 3500 m and 4000 m above sea level in altitude. This was a substantial barrier that blocked the moist Pacific atmosphere from reaching westward into the hinterland, resulting in the arid and hot semi-desertized climate in the Yunmeng Lake water system area.

1. INTRODUCTION

After the Triassic, a large-scale marine regression took place on the Chinese continent. The Jurassic and Cretaceous rocks of the hinterland are mainly nonmarine in origin. The lake systems in southwestern China were linked by the ancient Yangtze River, flowing east to west into the Tethys [1] giving rise to an independent fresh-water biogeographic province during

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the Cretaceous. The Tan-Lu Fault controlled the paleoenvironmentary changes during the Early Cretaceous in the volcanic belt of eastern China. The latest Jurassic and earliest Cretaceous volcanism was widespread in northern and southeastern China, and the fluvial and lacustrine deposits formed many intercalations within these volcanic sequences between the eruptions. They contain rich fossil plants and fresh-water animals such as well-known early birds *Confuciusornis, Liaoxiornis, Sinornis* and *Cathayornis*; feathered dinosaurs *Sinosauropteryx, Protachaeopteryx* and *Caudipteryx*; early angiosperm *Archaefructus* etc. [2, 3, 4, 5]. The inland water system of the Yunmeng and Songhua Lakes remained in northeastern and southeastern China, respectively, during the Late Cretaceous, and were characterized by different fossil faunas [6]. For example, the *Euestherites* conchostracan fauna in the Songhua Lake is composed of members of the Jilinestheriidae, Halysestheriidae, Estheriteidae and Dimorphostraciidae. Because of the temperate humid paleoclimate and their rapid evolution, this fauna diversified to include more than 100 species in 20 genera; the *Tenuestheria* conchostracan fauna to be restricted to the Yunmeng Lake system consists of members of the Sinoestheriidae, Euestheriidae, Afrograptidae and Paleolynceidae. It includes only about 15 species in 6 genera due to hot and arid contemporary climate as evidenced by the preponderance of red beds, gypsum and rock salt.

In this paper the author presents some accounts for the outlines of the paleogeography and environments in eastern China during the Cretaceous, mainly based on the the analyses of biological assemblages and lithologic features of strata in the latest Jurassic to early Early Cretaceous, late Early Cretaceous and Late Cretaceous times, respectively.

2. LATEST JURASSIC AND EARLY EARLY CRETACEOUS

The Latest Jurassic volcanism was widespread in North China, especially in the Great Hinggan Range, western Liaoning, northern Hebei and eastern Inner Mongolia. These volcanic rocks consist mainly of andesites, basalts, liparites, agglomerates, and tuffs. Many of the clastic sequences that were deposited in small and shallow lakes and swamps throughout this area between the eruptions are highly tuffaceous and contain rich and diverse assemblages of nonmarine bivalves, gastropods, ostracods, conchostracans, insects, shrimps, fishes, turtles, crocodiles, dinosaurs, pterosaurs, birds, mammals, charophytes and plants. They constitute the well-known Jehol biota [7].

In eastern Heilongjiang of northeastern China (Figure 1), the Wusuli Gulf, an arm of the Pacific Ocean, penetrated as far as the Suibin area. It was existent through the Early Cretaceous and contains intercalated volcanics.

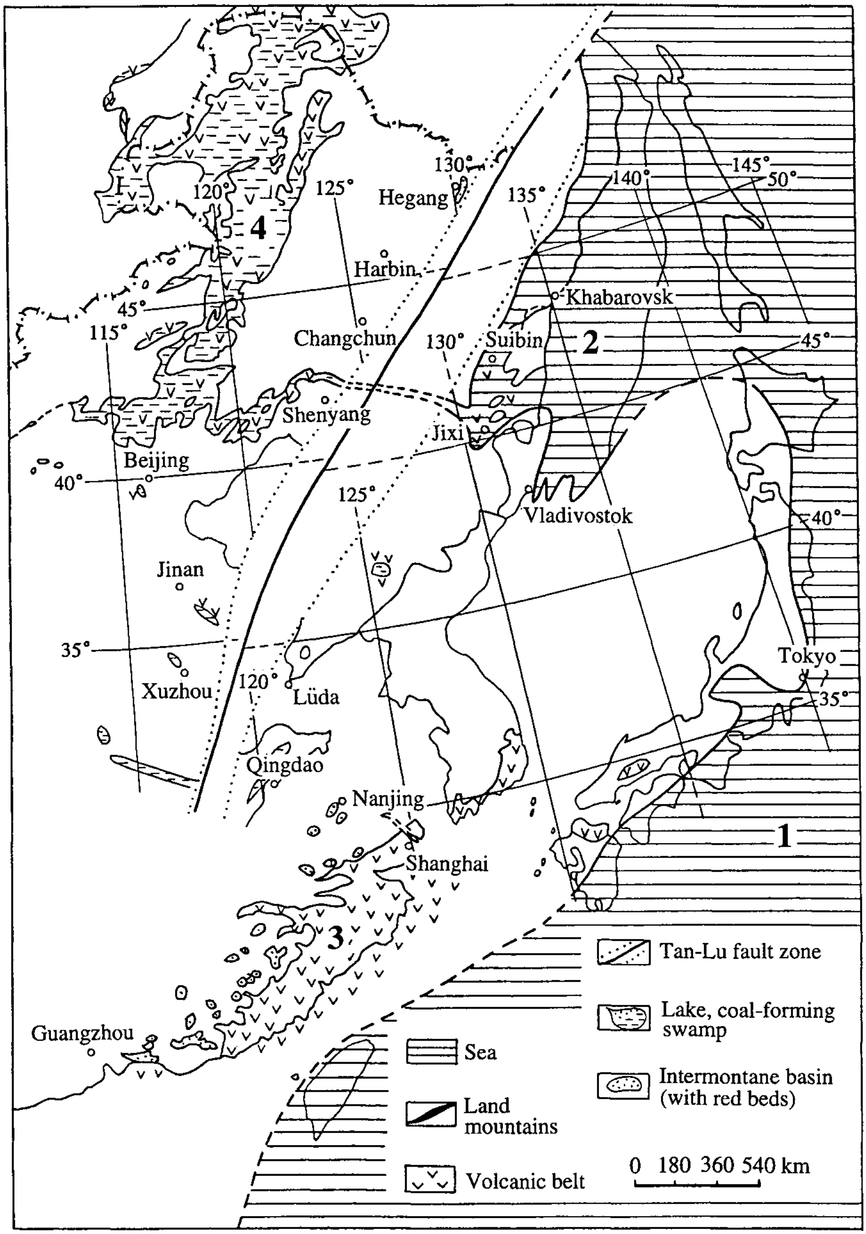

Figure 1. Sketch map showing the reconstruction of the latest Jurassic to early Early Cretaceous paleogeography of eastern China. 1: Pacific Ocean, 2: Wusuli Gulf, 3: active volcanic belt in the eastern coastal lowland, 4: volcano-swamp coal-bearing basins, 5: Songhua Lake, 6: Yunmeng Lake, 7: ancient Gan River, 8: southeast coastal mountains along the Pacific.

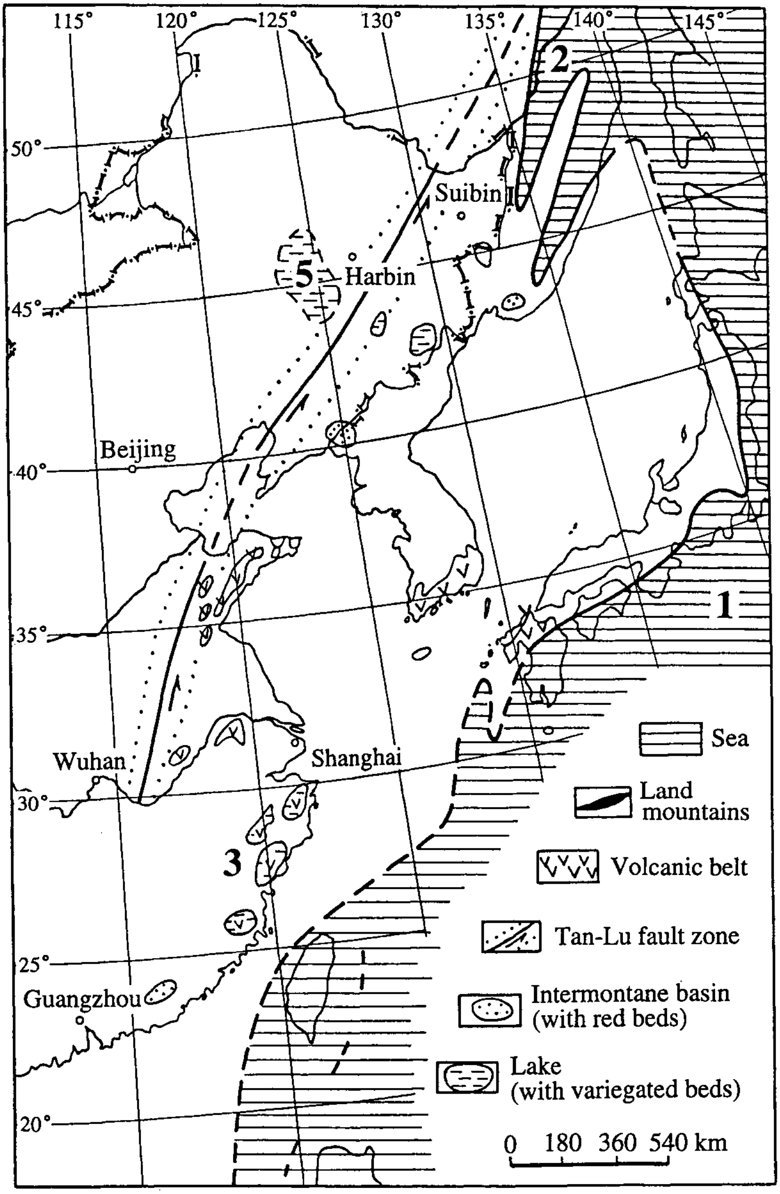

Figure 2. Sketch map showing the late Early Cretaceous paleogeography of eastern China. For further explanation of symbols see Figure 1.

Following the deposition of Lower and Middle Jurassic coal-bearing sequences in northern China, a major change in climate from humid to relatively arid took place during the Late Jurassic. This climatic change is indicated by widespread replacement of organic-rich, fluvial-lacustrine facies with nonmarine red beds. It has been suggested that this was caused by the restructuring of global atmospheric circulation patterns in response to the breakup of Pangea and an ensuing regional breakdown of monsoonal circulation [8]. This temperate and arid environment is indicated by the Jehol flora with narrow and tiny leaves and thick root system [9].

During the Early Cretaceous the climate of northeastern China returned to warm and humid with vast forests and many coal-bearing basins formed. In the meantime, an active volcanic belt was distributed in southeastern China extending into southern Korea and southwestern Japan along the coastal lowland of the West Pacific (Figure 1). From the sedimentary intercalations of Lower Cretaceous volcanic formations are found abundant fossil ostracods, conchostracans, bivalves, gastropods, insects, fish and plants, which indicate subtropical hot and arid climate [10].

3. LATE EARLY CRETACEOUS

The Tan-Lu fault zone, a large-scale shear fault trending north-northeast in eastern China, begins at Guangji of southeast Hubei in the south and extends northward through Lujiang in Anhui, Tancheng in Shandong, the Bohai Sea, Shulan in Jilin and Yilan in Heilongjiang before crossing the Chinese border into the Far Eastern Russian. Its total length within China is about 2400 km. It was active for a long period and controlled the lithology and stratigraphy along both sides of the fault from the Proterozoic up to the present time, but the large-scale horizontal displacement along the Tan-Lu fault zone started from the Aptian and ended within the Albian Stage with a horizontal strike-slip distance of 740 km. A seaway and possible river connected northeastern China with Japan, and that both the western Liaoning and Jixi areas of eastern Heilongjiang probably were geomorphologically similar and located at the same latitude from the latest Jurassic to the earliest Cretaceous (Figure 1). It was only after the large-scale horizontal displacement associated with the Tan-Lu fault zone that the distribution of late Early Cretaceous basins in eastern China occurred (Figure 2).

The Aptian marine transgression in the Taiwan-Penghu region is indicated by the ammonite *Holcophyllum caucasicum taiwanum* [11]. During this time there were only a few changes of the volcanic lowland along the southeastern coast of China, southern Korea and southwestern Japan, but the size of volcanic basins in southeastern China was much reduced

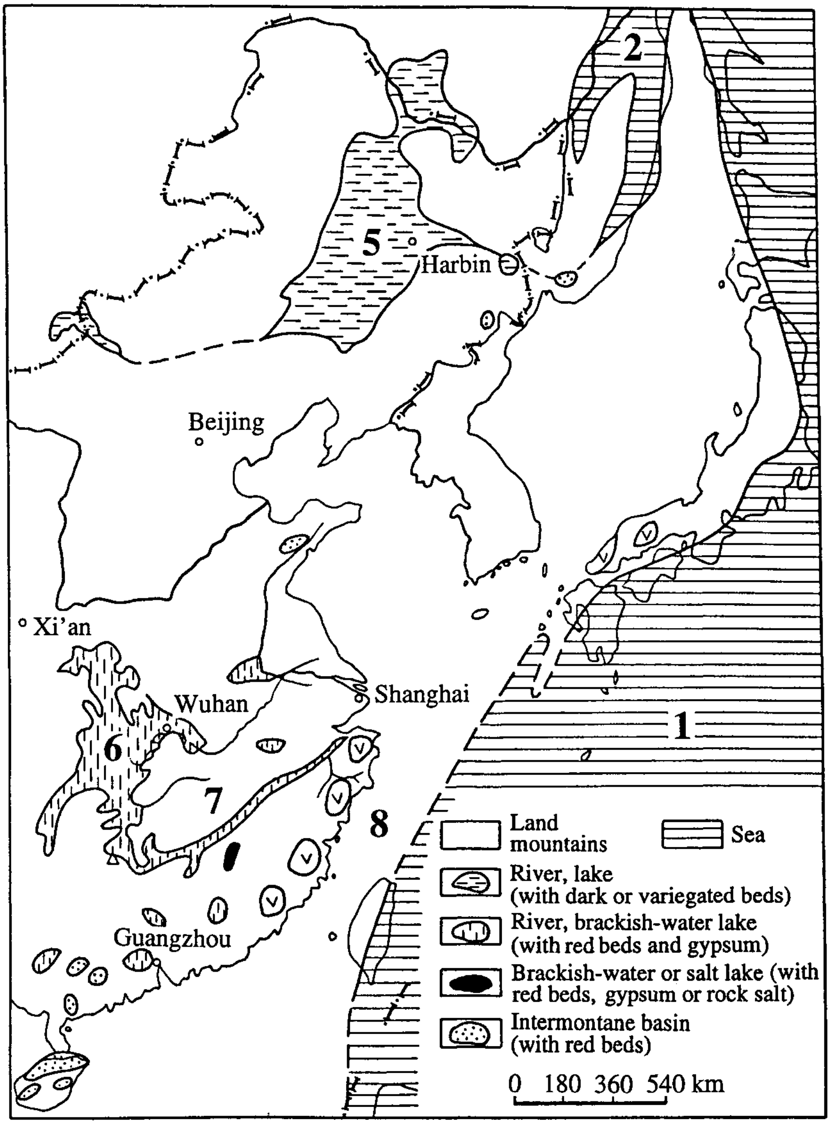

Figure 3. Sketch map showing the Late Cretaceous paleogeography of eartern China. For further explanation of symbols see Figure 1.

(Figure 2).

4. LATE CRETACEOUS

By the end of the Early Cretaceous widespread tectonic activity caused by the collision of the Himalayan and Lhasa Terranes [12], which dominated the depositional history of China and was called the Zhejiang-Fujian Movement in the southeastern coastal region, or the fourth episode of the Yanshan Movement elsewhere. It greatly altered paleogeographic patterns throughout China.

There were two newly formed huge lakes in the central China and Manchuria regions. The Yunmeng Lake occupied a large area of the southern Henan, Hubei, Hunan and a part of the Jiangxi Provinces. A major river system fed this lake from the Zhejiang and southeastern Jiangxi Provinces [1]. The Songhua Lake in Manchuria was connected to the north with the Bureying Lake in the Far East of Russia, with occasional influxes from the Pacific Ocean (Figure 3). This region contains an important source of petroleum.

During the Cenomanian-Santonian interval the ancient Pacific Ocean not only extended to near the northeastern corner of Heilongjiang Province from the Wusuli Gulf, but possibly also covered the eastern half of Taiwan. The latter area has yielded some Late Cretaceous corals (*Elephantaria*) from metamorphic rocks in the Bihou Group east of the Central Mountains.

In the early stage of Late Cretaceous the Lishui-Haifeng fracture began to be activated, due to the westward subduction of the Pacific Plate. East of this fracture, the eastern part of the Zhejiang-Fujian-Guangdong Provinces rapidly elevated to form the Coastal Mountains with the deposition of substantial thick-bedded molasse accumulations during the elevation on the western slope and the piedmont belt; these are called the Fangyan Formation or the Hutoushan Conglomerate in Zhejiang; the Chishi Formation or the Upper Subgroup of the Shimaoshan Group in Fujian, and the Danxia Formation in northern Guangdong (Figure 4). All are composed of massive sandstone and conglomerate, ranging from several hundred meters to 1735 m, with a maximum thickness possibly exceeding 2000 m. After longstanding weathering and denudation, the principal remnants are currently distributed in the Yongkang, Jin-Qu, Ningbo, Xiakou and Si'an Basins of Zhejiang; the Shaxian, Shanghan, Cong'an and Hekou Basins of Fujian, and the Pingshi and Danxia Basins, together with those small basins around the Lianxian-Yangshan area in Guangdong. In the small-sized intermontane basins east of the Lishui-Haifeng fracture, there are intermediate to acidic volcanic rocks intercalated with the conglomerate beds. These strata are called the Tangshang Formation in eastern Zhejiang, and the Upper Subgroup of the Shimaoshan Group in Fujian. In eastern Guangdong, they very

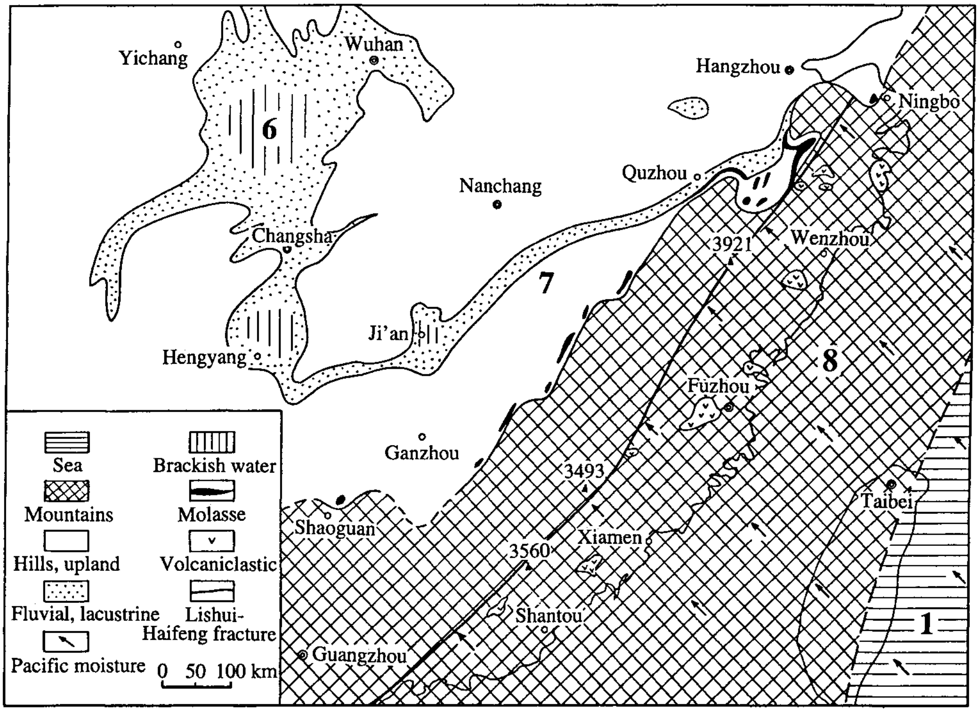

Figure 4. Sketch map showing the Late Cretaceous litho-paleogeography of central and southeast China. For further explanation of symbols see Figure 1.

likely correspond to a pink rhyolite intercalated between the lower and upper red beds in the Xingning Basin. From the mudstone lenticulations intercalated in these great thick-bedded sandy conglomerates, occasionally can be found fossil conchostracans, ostracodes, fish, dinosaurs, sporopollen and plants of lower Upper Cretaceous affinity, such as *Linhaiella, Feiyunella, Yumenella, Paraclupea, Zhejiangopterus linhaiensis, Frenelopsis parceramosa, Pagiophyllum* sp. and *Cupressinocladus elegans,* and a sporopollen assemblage dominated by *Schizaeoisporites* with a small amount of angiospermous pollen.

Based on the mean altitude of the modern Zhejiang-Fujian mountainous area combined with the thickness of the denuded molasse accumulations, the Coastal Mountains of southeast China during the Late Cretaceous ranged between 3500 m and 4000 m above sea level in altitude, with a width of about 500 km from east to west [13]. Such a huge barrier was much more than the Qinling-Dabieshan Mountains, the modern geographical and climatic boundary between North and South China. This barrier completely blocked the moist Pacific atmosphere from reaching westward into the hinterland (Figure 4); as a result, the vast plain and hilly area of the Yunmeng Lake water system in Central China began to become a tropic (subtropic) arid and hot semidesertized and salinized lake district. This can be well confirmed by the Upper Cretaceous red clastic deposits and the eolian sand and wind-scoured dreikanters contained therein, together with gypsum in the Jiang-Han and Hengyang Basins, in addition to the rock salt deposited in local places (as in Zhoutian, Jiangxi). Up to the Eocene, as a result from the further development of desertization and salinization of the lakes, relatively thick gypsum and rock salt were extensively formed in the Jiang-Han, Hengyang and some other basins.

5. CONCLUSION

In this paper, the author has described only the paleogeographical patterns and paleoenvironmental changes during the latest Jurassic to Cretaceous in eastern China, but the origin and evolution of the basins in this region might have been mainly caused by the large-scale plume-related magmatisms and strike-slip faults in the eastern margin of the Asian continent, as Okada [14] has pointed out recently.

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Upper Mesozoic unconformity-bounded units of Korean Peninsula within Koguryo Magmatic Province

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The Upper Mesozoic strata of the Korean Peninsula are subdivisible into three unconformity-bounded units, and they developed time-transgressively toward the Pacific. The lower unit straddling the Jurassic-Cretaceous boundary occurs profusely in Northern Korea with substantial amount of volcanic rocks but is extremely rare in Southern Korea. The middle, Hauterivian-lower Albian, unit occurs scattered in the Korean Peninsula but preponderantly in the Kyongsang Basin, Southeastern Korea. The upper unit is extremely rare in North Korea but is abundant in the southern part of Southern Korea, where it is characterized by calc-alkaline volcanic rocks. Such a tendency of southeastward younging is also apparent over the 'Koguryo Magmatic Province' inclusive of Korea, suggesting that the shifting was caused by the retreating subduction zone on the Pacific side. But the post-Triassic magmatisms were largely keeping the same overall area suggesting a sustained hot spot above a plume below. The association in space and time of magmatisms and sedimentations implies their genetic relation. The Cretaceous sedimentary basins of Southern Korea are interpreted as pull-apart origins under a strike-slip tectonics.

1. INTRODUCTION

In the Korean Peninsula, the Cretaceous System is physically inseparable from a natural Upper Jurassic-Paleocene unit with bounding unconformities. Though the stratigraphic data yet available from Northern and Southern Korea are still meager, they proved it possible that the Upper Jurassic-Paleocene unit is sudivided into three unconformity-bounded

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Some of the North Korean data of this paper derive from a paper co-authored with Prof. Nadezhda Filatova, Institute of Lithosphere of Russian Academy of Sciences, whom we wish to express our thanks. Special gratitudes are due to Dr. Niall J. Mateer of the University of California who reviewed and distinctly improved the manuscript.

units. In the process of describing these units, their distributions over the Korean Peninsula have been found to show a distinct tendency of shifting with time toward the Pacific. Beyond this comparative stratigraphy of Northern and Southern Korea, we have reached a new horizon for the Korea's geologic setting in East Asia.

For the North Korean data we had to rely on publications from North Korea. As to the South Korean data, we cited the Kyongsang Basin, Southeast Korea, too frequently due both to the superior records of the basin and the authors' disciplines. The field experiences acquired from the Kyongsang and other basins in South Korea are supposed to be effective for understanding all the other sedimentary basins dealt with in this paper.

The most impressive was the development of the sedimentary basins and volcanic zones showing a distinct younging toward the Pacific. To understand such southeastward younging in a wider geographic extent beyond the Korean Peninsula, a review has been made of geologic maps [1, 2, 3, 4] and informations. As an outcome of such a pursue, a more comprehensive, subcircular shaped, post-Triassic geologic province here called the 'Koguryo Magmatic Province' has been recognized, the significance of which is discussed in the later part of this paper.

2. UNCONFORMITY AND TECTONICS

The Mesozoic East Asia including Korea was characterized by crustal upheaval and nonmarine sedimentations. After the Triassic collision of the Yangzi (Yangtse) against the Sino-Korea continents, the Jurassic Korea in the eastern continental margin of Asia underwent a post-collisional Daebo (Taebo) Orogeny, which was a crustal readjustment under the influence of the subduction from the Pacific side. The Daebo Orogeny was practically over in the Late Jurassic, when a new extension-dominant tectonic regime was opened for Korea and the vicinity. Thus, the Mesozoic Erathem in Korea and the vicinity would be tectono-stratigraphically divided into the Lower Mesozoic and the Upper Mesozoic, the divide being approximately the boundary between the Middle Jurassic and the Upper Jurassic [5, 6, 7, 8].

The Kyongsang Supergroup of the Kyongsang Basin and their correlatives over the Korean Peninsula are collectively called the Kyongsang Synthem [9, 10, 11, 12]. Major unconformities and tectono-stratigraphic considerations divide the Upper Mesozoic strata in Korea into the Jasong and the Kyongsang Synthems and the latter into two subsynthems: the Lower Kyongsang Subsynthem and the Upper Kyongsang Subsynthem. Thus, we have in Korea three unconformity-bounded units of the Upper Mesozoic.

The bounding unconformities for these three units reflect four diastrophic phases: the mid-Jurassic phase of the Daebo Orogeny, Valanginian(-Hauterivian) Nakdong Disturbance [30] (in North Korea, it is called the Jaeryonggang Tectonic Movement [14]), mid-Albian Yuchon

Disturbance, and the last phase of the Bulguksa Disturbance that marks the end of the Kyongsang Basin.

The Jurassic-Cretaceous boundary in Northern Korea, like that in NE China, occurs within a sedimentary continuum. Thus, the stratigraphic unit that straddles the boundary is called the Jasong System in Northern Korea [13, 14], which is called here the Jasong Synthem as it is a major unconformity-bounded unit. The Valanginian(-Hauterivian) disturbance not only folded the Jasong Synthem but also affected a migration of its depocenter from Northern Korea to the Kyongsang Basin, Southeast Korea.

The sub-Yuchon unconformity [15] denotes a mid-Albian disturbance, here called the Yuchon Disturbance. The unconformity is obscure in some sections of the central part of the Kyongsang Basin but locally exhibit a spectacular angular discordance in such a place as Mt. Kumsong. In the periphery of the Kyongsang Basin, the Yuchon Group overlies either the Sindong Group or the basement exhibiting a considerable hiatus. Though the unconformity represents only a brief erosion and the disturbance only mildly folded the earlier strata, the unconformity represents a transition from the clastics-dominant to volcanics-dominant depositions and marks an abrupt widening and shifting of the depositional sites. Such a start of vigorous volcanism was soon succeeded by a petrochemical change from alkaline to calc-alkaline compositions [18] implying water supplies through subductions for the magma genesis.

The Aptian-lower Albian Hayang Group comprises some basalts and andesites which are alkaline to subalkaline [16, 17, 18] while the upper Albian-lower Upper Cretaceous Jusasan Andesitic Subgroup (the lower part of the Yuchon Volcanic Group) reportedly is partly alkaline and partly calc-alkaline; and the rest of the Yuchon Volcanic Group, mainly of acidic volcanic rocks, is distinctly calc-alkaline [18]. As reviewed below, the Hayang-Yuchon transition represents the so-called 'mid-Cretaceous tectonism', which is recognized interregionally.

The mid-Cretaceous (110-85 Ma; Albian-Coniacian) 'pulse' in the rate of sea-floor spreading [19] and the 'mid-Cretaceous' (Aptian-Albian: 124-97 Ma) 'pulses' of peri-Pacific deformation, metamorphism, uplift and hiatus in sedimentation [20] are under the same context with the mid-Albian unconformity in Korea. According to Kirillova [21, 22], the mid-Albian is the time of tectonism, volcanism, unconformity and transgression in NE Russia, and marked the initiation of the upper Albian-Senonian Okhotsk-Chukotka continental marginal volcanic belt. Filatova [23, 24] concluded that the interval of 120-100 Ma (mid-Cretaceous: Aptian-Albian) is the period of intense orogeny over the peri-Pacific and Tethyan continental margins due to collisions caused by an accelerated rate of the spreading, which in turn was caused by a Pacific superplume. She also concluded that the succeeding interval of 100-80 Ma (the early Late Cretaceous) is an anomalously intensive volcanism due to normalized subduction after the brief cessation of the subduction.

But, in Korea, there was only a mild deformation, instead of an orogeny, during the Aptian-Albian interval. According to a study on plate motions in the Pacific Basin relative to the surrounding continents [25], the Izanagi plate was spreading in the Aptian-Albian toward the north almost parallel with the East Asian continental margin. If it was the case, a subduction and the related compressional tectonics could not be expected. The alkaline volcanism of the Aptian-Albian interval and the calc-alkaline one of the enormous quantity in the Cenomanian and the succeeding Late Cretaceous correspond, respectively, to an intra-plate volcanism during the cessation of the subduction in the Aptian-Albian and a series of intense volcanism caused by speeded subduction when it was resumed after a cessation.

The Kyongsang Supergroup, the stratotype of the Kyongsang Synthem, now appears to have continued to develop even in the lowermost Cenozoic according to recent isotope datings, which mean that the Late Cretaceous volcanism slightly surpassed the K-T boundary to finish up in the Paleocene(-Eocene), a tectono-stratigraphic boundary that serves the top of the unconformity-bounded unit.

3. STRATIGRAPHIC DESCRIPTION

3.1. Jasong Synthem

This Upper Jurassic-lowermost Cretaceous unit is characterized in Northern Korea by volcanisms showing three stages of a compositional cycle typical to the development of a volcanic belt: (1) clastic-pyroclastic deposits, (2) thick volcanic association showing various compositional differentiation and (3) alternated sedimentary and pyroclastic rocks [26].

The Jasong Synthem occurs in the following major localities in Northern Korea:

3.1.1. Daedonggang Basin

Strata of the Daedonggang Basin near Pyongyang (Figure 1, loc. 4) are about 1550 m thick, subdivided into the Chimchon, Daebosan and Mangyongdae Groups (in ascending order) (Figure 2).

The Chimchon Group, about 300-400 m thick, is composed of conglomerates, sandstones, siltstones, tuffaceous sandstones and tuffs. The tuffs of intermediate composition predominate here with basaltic, andesitic, and andesite-dacitic pyroclasts ranging in size several mm to 2-3 cm. Subordinately, acidic tuffs also occur. Filatova identified in this unit some mylonitized clasts of the Lower Jurassic Songnimsan Series, which indicate a tectonism before the accumulation of the Chimchon Group [26].

The Daebosan Group is a volcanic association, about 800 m thick. Near Pyongyang City, 5-10 dark-gray andesitic lava flows (130-170 m thick) occur at its base. Overlying are intercalated medium- to coarse-grained tuff layers mainly of andesitic composition with subordinate amount of dacitic tuffs. The upper part of the Daebosan Group consists of the rhyolitic lava-flows and numerous extrusive domes with rhyolitic breccia.

The Mangyongdae Group, about 350 m, consists of tuffaceous sandstones, acidic tuffs and conglomerates.

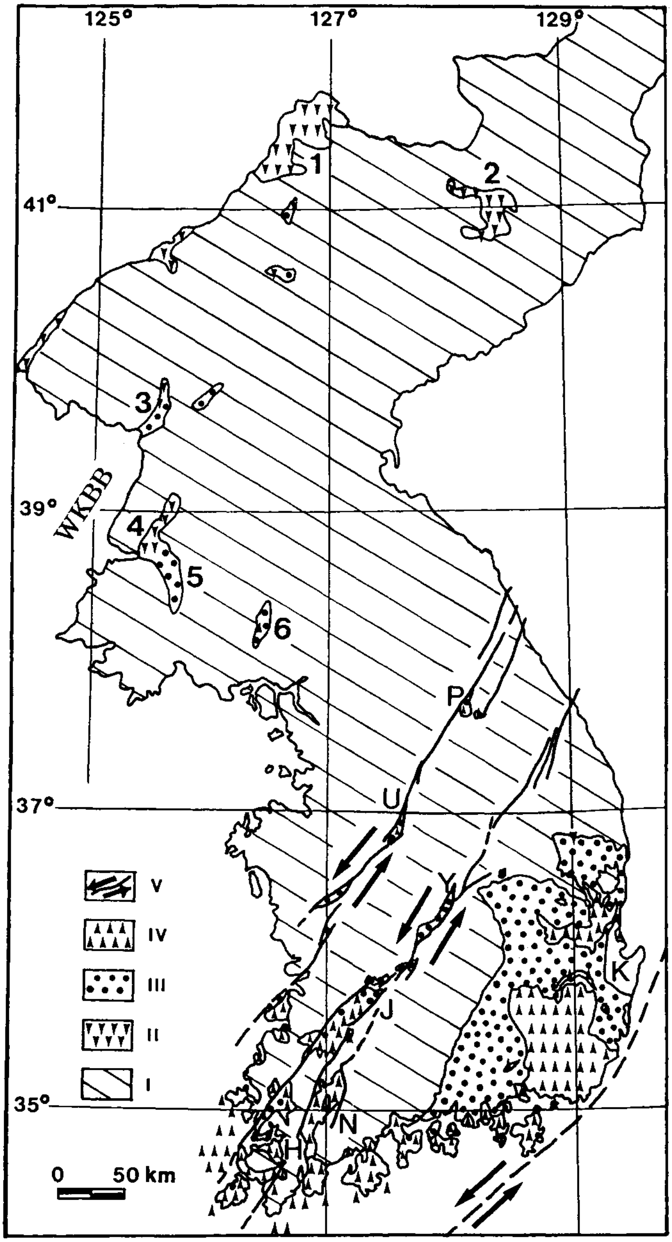

Figure 1. Upper Mesozoic sedimentary basins of Korean Peninsula showing basin-fills by age.
I: basement (Precambrian-Jurassic); II: Upper Jurassic-lowermost Cretaceous Jasong Synthem; III: Hauterivian-lower Albian Lower Kyongsang Subsynthem; IV: upper Albian-Cretaceous Upper Kyongsang Subsynthem; V: Cretaceous faults with arrows showing sinistral motions.
K: Kyongsang Basin,
Y: Yongdong Basin,
J: Jinan Basin,
N: Nungju Basin,
H: Haenam Basin,
P: Pungam Basin,
U: Umsong Basin.
1: Amnokgang Basin,
2: Hochongang Basin,
3: Daeryonggang-Chongchongang Basin,
4: Daedonggang Basin,
5: Jaeryonggang Basin,
6: Yesonggang Basin,
WKBB: West Korea Bay Basin.

3.1.2. Jaeryonggang Basin

In the Jaeryonggang Basin (Figure 1, loc. 5), the Chimchon Group is missing, and the Daebosan Group composed of lavas and tuffs rests unconformably on the Lower Jurassic Songnimsan Series (Figure 2). The Daebosan Group here includes a volcanic association of andesite, dacite, basalt and rhyolite. The bomb tuffs of basaltic and andesitic compositions prevail in the basal part, and the lapilli of dacitic composition are widespread in the top of the sequence. The youngest, acidic, shallow intrusive bodies and extrusive domes occupy the central and peripheric parts of the cauldrons. The volcanic squence of the Daebosan Group here is thus similar with that of the same group in the Pyongyang area. The overlying tuffaceous-sedimentary sequence, a correlative of the Mangyongdae Group, has a very limited distribution.

3.1.3. Daeryonggang Basin

In the Daeryonggang Basin along the Daeryonggang fault zone near Anju City (Figure 1, loc. 3), the Bongsu Group, 600-900 m thick, contains distinctly mafic volcanic rocks; it is an undivided equivalent of the Chimchon, Daebosan and Mangyongdae Groups. The lower part of the Bongsu Group consists of conglomerates, sandstones and mudstones with pyroclastic admixtures. The middle part is composed of dark-green and gray violet basalts and andesitic basalts as well as basic tuffs. The basalts contain the phenocrysts of plagioclase, olivine, clinopyroxene, rare hornblende and ore minerals. The upper part of the group consists of tuffaceous-sedimentary rocks including acidic tuff layers.

3.1.4. Amnokgang Basin

In the Amnokgang Basin near Jasong City (Figure 1, loc. 1), the Yonmuri Group, about 2000 m thick, is correlated with the Sinuiju Group, and subdivided into the Changpyong, Tochang and Chonjaebong Formations in ascending order. The former two formations consist mainly of intermediate to acidic volcanic rocks but include variegated and gray clastics and thin coal beds. The lower part of the Chonjaebong Formation consists of trachyte, trachytic tuff and felsic volcanic rocks, and the upper part consists of quartz porphyry, felsite and their tuffs. Thus, the volcanic sequence of the Yonmuri Group has the early phase of the intermediate-acidic compositions and the last phase of sub-alkaline felsic composition [14, 18, 26].

3.1.5. Hochongang Basin

In the Hochongang Basin near Kapsan City (Figure 1, loc. 2) an analogous volcanic sequence of andesites, dacites and rhyolites occurs.

3.1.6. West Korea Bay Basin

The West Korea Bay Basin is an offshore sedimentary basin located near 39th N. Latitude (Figure 1). The maximum drilled sequence, 750 m thick, is divisible into the lower (fluvial siltstone, shale, sandstone and thin coals; 200 m thick) and the upper (fluvio-lacustrine black shale, sandstone, thin limestone beds and basal conglomerate; 550 m thick). The upper unit, correlated with the Shinuiju Formation, was deposited in a large, deep freshwater tectonic lake under a humid climate and contains some possible oil source rocks [27].

In Southern Korea, the Jasong Synthem is represented by the Myogok Formation that occurs in a very limited northern peripheral area of the Kyongsang Basin. The formation, about 70 m thick, is composed of sandstone and black shale of the fresh-water facies, without volcanic layers. The geologic age of the formation has been assigned either the Late Jurassic based on fresh-water molluscs [28] or as Tithonian or possibly the earliest Cretaceous based on palynomorphs [29]. The strata are folded by the Nakdong Disturbance (Valanginian) [30].

3.2. Lower Kyongsang Subsynthem

The Kyongsang Synthem is here divided into the Lower and the Upper (see Chapter 2). The stratotype of the Lower Kyongsang Subsynthem is the Sindong and the Hayang Groups together of the Kyongsang Basin. These two groups and the overlying Yuchon Volcanic Groups (Table 1) constitute the Kyongsang Supergroup, the stratotype of the Kyongsang Synthem of the Korean Peninsula (Figure 2). A major Late Mesozoic sedimentary basin located in Southeast Korea, the Kyongsang Basin developed under fluvio-lacustrine conditions throughout its history. In this comparative study of the Northern and Southern Korean Cretaceous basins, the Kyongsang Basin represents, as it deserves so, all coeval sedimentary basins of Southern Korea. The Cretaceous and related stratigraphic classification of this paper has been grown from the past studies of the Kyongsang Basin and other coeval basins of Southern Korea [34, 35, 36, 10, 11, 37, 38] and reinforced by the informations on the Northern Korean Upper Mesozoic basins [14, 26].

In the Hauterivian-Barremian time interval, the NNE trending basin called the Nakdong Trough was extant now occupying the western part of the Kyongsang Basin, in which the Sindong Group, 2000 to 3000 m thick, composed of sandstone, shale, conglomerate and marl was deposited. The major source area to the WNW (the Lower Mesozoic Okchon Orogen), as indicated by paleocurrent analysis, was uplifted intermittently as reflected in sedimentary megacycles [32]. A rapid basinal expansion toward E to NE in the Aptian brought to the deposition of the Hayang Group over the whole Kyongsang Basin. This group, 1000 to 5000 m thick, is composed of shale, sandstone, conglomerate and contemporaneous volcanic rocks. Paleocurrent and facies analyses indicate a dominant clastic source area to the east, the Permian-Jurassic accretionary complexes of Southwest Japan, which was then adjacent to the Kyongsang Basin [31, 32, 33, 39].

During the sedimentation of the Hayang Group, the Kyongsang Basin was dominated by WNW-trending growth faults that split the basement. Syndepositional block movements and related sedimentation yielded abrupt lateral lithologic and thickness changes causing difficulties in stratal correlation within the Kyongsang Basin [9, 12, 33]. The Sindong and Hayang

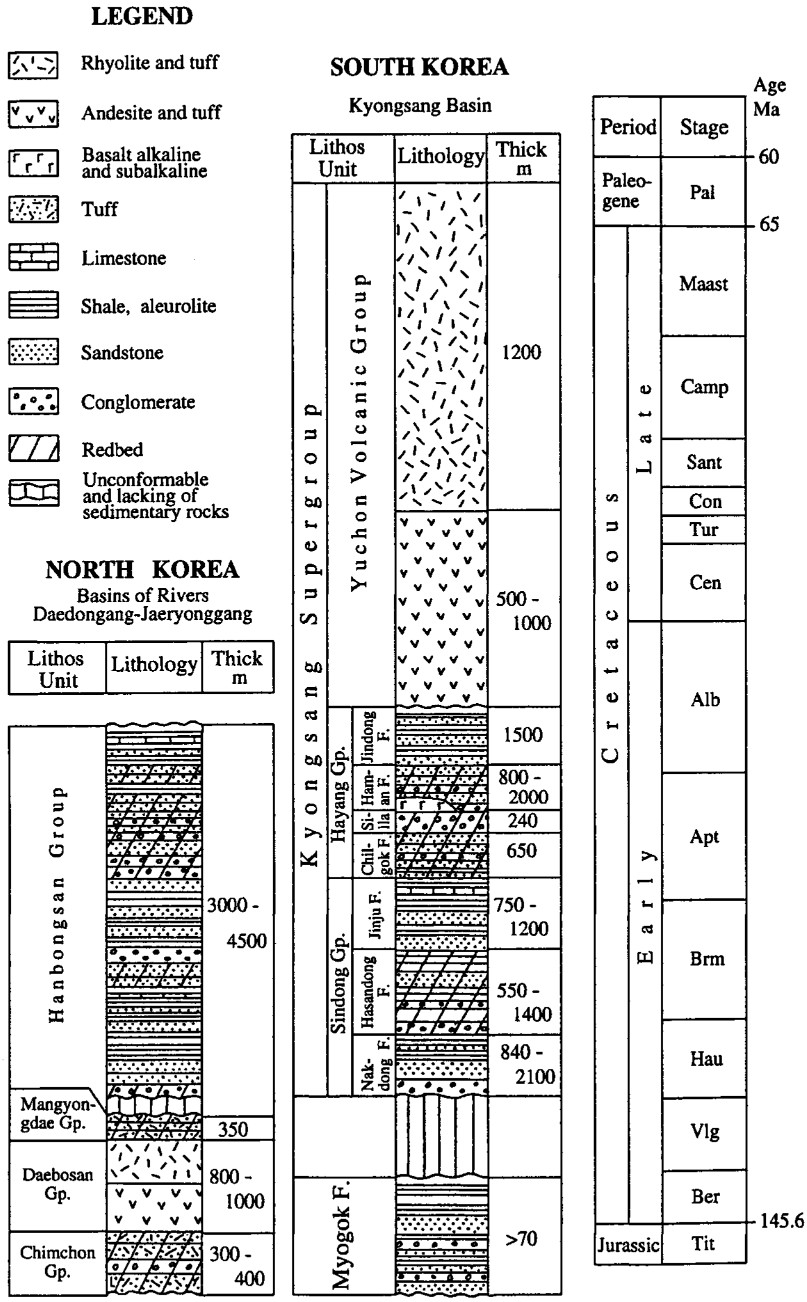

Figure 2. Highly simplified geological columnar sections representative of basins in northern and southern parts of Korean Peninsula.

Groups form a single sedimentary cycle in the northern part of the Kyongsang Basin, while they form two distinct sedimentary cycles in the southern part of the same basin; this contrast suggests that the sedimentary cycles here were determined mainly by local tectonics; a unified sedimentary cycle throughout the Korean Peninsula may have existed, but seems to have been extensively disturbed by local vertical tectonics of crustal blocks.

The volcanic pebbles of the Hayang Group, derived from contemporaneous volcanic sources, contrast with the pebble composition of the Sindong Group, in which no volcanic pebbles are observed. Such a stratigraphic significance of the pebble compositions may be only locally applicable within a single sedimentary basin. In the northern part of the Kyongsang Basin, the Hayang Group yields unique conglomerates of the Kumidong and Kisadong Formations comprising the pebbles of the Permian, Triassic, Jurassic and Lower Cretaceous cherts. In lithology and fossil content, these cherts are typical of the Tamba-Mino-Ashio and Chichibu belts of Japan [36].

In about the beginning of the Aptian, an abrupt eastward basin migration took place, i.e., from the Nakdong Trough to the whole Kyongsang Basin. Such a peculiar basin development awaits a technical study which may explain with a pull-apart model; the southern part of the Korean Peninsula was sliced by several sinistral strike-slip faults, one of which underlies the Korea Strait, the Korea-Taiwan Strait Fault [37]. The same geodynamic system might have formed various Cretaceous sedimentary basins in Southern Korea including the Kyongsang Basin.

Along the Okchon-Taebaeksan Zone, the strata belonging to this unit were accumulated in several elongated pull-apart basins created by the sinistral strike-slip movements of the Yongnam, Okchon-Taebaeksan, and Kyonggi blocks during the Cretaceous Period. Most of the basins, bounded by one or two strike-slip faults, were formed along the both sides of the block [38, 39, 40, 41, 42, 43, 44, 45, 46, 47]. These papers report that the basins started to form not coevally but at varying times during the Cretaceous Period; most of them were formed during the time of the Hayang sedimentation but some of them formed earlier or later (Table 1). The basin-forming strike-slip faults served as the outlet of coeval volcanism during sedimentation forming an intertonguing facies of sedimentary and volcanic layers.

Northern Korean Cretaceous strata that unconformably overlie the Jasong Synthem are characterized by a lack of a contemporaneous volcanic component except some reworked volcanic debris of acidic and andesitic compositions derived from the eroded Jasong Synthem [14]. The Hauterivian-lower Albian unit of Northern Korea yielded *Clypeator jiuquanensis*, a Hauterivian-Barremian charophyte, and a flora that characterizes the upper part of the Lower Cretaceous [14].

3.2.1. Daedonggang-Jaeryonggang Basin

In the Daedonggang-Jaeryonggang Basin, the Hanbongsan Group, 3000-4000 m thick, unconformably rests on the Upper Jurassic-Neocomian strata and contains numerous volcanic

Table 1
Correlation of Upper Mesozoic stratigraphic units of Korean Peninsula. Some guide taxa are shown: *Trigo.: Trigonioides; Plicat.: Plicatounio; Eoses.: Eosestheria; Nemes.: Nemestheria;* and *Neodi.: Neodiestheria*

<table>
<tr><th rowspan="2">Region \ Age</th><th rowspan="2">whole KOREA</th><th colspan="9">NORTH KOREA</th><th colspan="7">SOUTH KOREA</th><th rowspan="2">whole KOREA</th></tr>
<tr><th colspan="3">Amnokgang Basin</th><th colspan="2">Daeryonggang-Chongchongang Basin</th><th>Daedong-gang-Jaeryong-gang Basin</th><th>Yesong-gang Basin</th><th>Duman-gang Basin</th><th>Yong-dong Basin</th><th>Jinan Basin</th><th>Nungju Basin</th><th>Haenam Basin</th><th>Pungam Basin</th><th>Umsong Basin</th><th>Kyong-sang Basin</th></tr>
<tr><td>K2-Paleoc.</td><td>Upper Kyongsang Sub-synthem</td><td colspan="3"></td><td colspan="2"></td><td></td><td>Bonghwa-san Group</td><td></td><td>volc. rocks</td><td>volc. rocks</td><td rowspan="2">Nungju Group</td><td>Haenam Group</td><td>volc. rocks</td><td>volcan. rocks</td><td>Yuchon Volc. Group</td><td rowspan="3">Kyong-sang Synthem</td></tr>
<tr><td rowspan="2">K1</td><td rowspan="2">Lower Kyongsang Sub-synthem</td><td>Sujin Group</td><td colspan="2">Bongchon-bong Group</td><td rowspan="2">Bakchon Group
Trigo.
Plicat.</td><td>Gunuri Group</td><td>Hanbongsan Group
Nemes.</td><td>Sansongri Group
Pyonghwa-ri Group</td><td rowspan="2">*Neodi.*
Seson Group
Trigo.
Plicat.</td><td rowspan="2">Yong-dong Group</td><td>Jinan Group</td><td></td><td>Pungam Group</td><td rowspan="2">Umsong Group</td><td>Hayang Group</td></tr>
<tr><td></td><td colspan="2"></td><td></td><td></td><td></td><td></td><td></td><td></td><td></td><td>Sindong Group</td></tr>
<tr><td>J3</td><td>Jasong Synthem</td><td>Sinuiju Group
Eoses.</td><td>Yon-muri Group</td><td>Yong-song Group</td><td>Bongsu Group</td><td>Sohwari Group</td><td>Mangyong-dae Group
Daebosan Group
Chimchon Group</td><td></td><td>Yong-nam Group</td><td></td><td></td><td></td><td></td><td></td><td></td><td>Myogok Fm.</td><td>Jasong Synthem</td></tr>
</table>

fragments derived from the Daebosan and Mangyongdae Groups. The lower part of the Hanbongsan Group consists of red and gray conglomerates, sandstones, mudstones with limestone layers. The middle part includes conglomerates below and alternation of sandstones, siltstones and mudstones above. The upper part is composed of conglomerates, sandstones and mudstones below and red, green and gray sandstones, siltstones, mudstones with limestone layers above (Figure 2).

3.2.2. Daeryonggang Basin

In the Daeryonggang Basin, unconformably overlying the Bongsu Group, the Bakchon Group, about 2000-2500 m thick, comprises red conglomerates that pass above into alternated conglomerates, sandstones and siltstones of the upper part intercalated with gray and black argillites, siltstones, sandstones and rarely limestones.

3.2.3. Amnokgang Basin

In the Amnokgang Basin, the Bongchonbong Group, about 1800 m thick, comprises thick basal conglomerates (about 300 m) and two levels of conglomerates in the middle and upper parts. Black siltstones and fine sandstones occupy two levels: above the basal conglomerate and at the top. Sedimentary cycles are similar with those of the Daedonggang-Jaeryonggang Basin.

3.3. Upper Kyongsang Subsynthem

The stratotype is the Yuchon Volcanic Group, the upper group of the Kyongsang Supergroup. The Yuchon Group consists of volcanic and some clastic rocks, and unconformably overlies the Hayang Group and the earlier rocks [15]. The Group is subdivided into an andesitic lower part (Jusasan Intermediate-Volcanic Subgroup) which is partly alkaline and partly calc-alkaline and an acidic upper part (Unmunsa Acidic-Volcanic Subgroup) which is calc-alkaline. The former corresponds to the upper Albian-Cenomanian peak of the isotope ages of the Kyongsang Basin. Another more distinct Campanian-lower Maastrichtian peak of the isotope ages appears to indicate the culmination of the Unmunsa volcanism [48, 49]. The top of the Yuchon Volcanic Group may go up till the basal Paleogene in some sections according to isotope datings [18]. The Yuchon volcanics except the basal part and the comagmatic plutonic rocks are distinctly calc-alkaline, suggesting its association with the subduction-related continental marginal volcanic belt [17].

The group is particularly thick and widespread in the southern part of the Kyongsang Basin, and the sedimentary volume increases toward the south finally to dominate the volcanic layers. The group laterally extends toward the west beyond the basin, so that the correlatives of the group are widespread in the southern part of Southern Korea, particularly along the Korea Strait.

In the basins along the Okchon-Taebaeksan Zone, the volcanic-sedimentary sequences that are correlated with the Yuchon Group unconformably overlie either the Hayang Group equivalents, the Sindong Group equivalents or the Precambrian-Jurassic basement. In the Haenam Basin in the SW extremity of Korea, the Haenam Group, with a maximum thickness of 2.6 km, disconformably overlies the Jurassic granite. The lower unit, the Uhangri Formation consists of volcaniclastic conglomerate, sandstone and siltstone formed in lacustrine delta. The upper unit, the Hwangsan Tuff, consists of acidic tuff and conglomerate and sandstone. The group, coarsening-to-fining upward sequences of pyroclastics and epiclastics, was formed in semi-arid, alkaline and euxinic conditions in which a large amount of andesitic and rhyolitic ashes were supplied [40].

In Northern Korea, the Bonghwasan Formation in the Yesonggang graben is the only Upper Cretaceous unit in Northern Korea. The formation, about 424 m thick, consists of sandstone, mudstone and rarely conglomerate; the basal part, about 80 m thick, comprises gray fine clastics while the above comprises reddish coarser clastics. The Late Cretaceous age is based on plant fossils [13, 14, 50]. The Formation overlies the unfossiliferous Sansongri Formation reportedly conformably [14].

4. GEOCHRONOLOGY AND ENVIRONMENT

The Northern Korean strata of the Jasong Synthem yield abundant Late Jurassic-Neocomian plant fossils [51,52]. The Sinuiju Group yielded *Eosestheria middendorfii-Ephemeropsis trisetalis* (insect)-*Lycoptera davidii* (fish), which was known from the uppermost Jurassic Fushin Formation of NE China. The assemblage is also known from the Upper Jurassic of Mongolia and Trans-Baikal and from the Lower Cretaceous of Mongolia and Russian Primorye [53, 54].

The only reliable isotope age, so far available, of the volcanic rock that belongs to the Jasong Synthem is the K-Ar age of a trachyte body on Mt. Oknyobong, near Unsong of the Mungyong area, Southern Korea. One sample has shown 139 Ma, that suggests the earliest Cretaceous volcanism. The Northern Korean Amnokgang Intrusives that intruded the Upper Jurassic volcanic rocks of the Jasong Synthem yielded K-Ar ages, from the Jasong area ranging from 156 to 97 Ma (Late Jurassic-Early Cretaceous), and from the Junggang area 142 Ma (earliest Cretaceous) [14]. These ages suggest that they intruded in an interval that astrides the Jurassic-Cretaceous boundary.

The Upper Jurassic-lowermost Cretaceous Jasong Synthem is the correlative of the Jehol Group in NE Chian. In the Korean Peninsula, it occurs mainly in Northern Korea, characterized by volcanic rocks in it. The loci of major sedimentation apparently migrated in time toward southeast to unfold the sedimentation in the Kyongsang Basin in about the Hauterivian Age. Late Cretaceous strata occur maily in Southern Korea, which are volcanic and volcaniclastic rocks. The vigorous mid- to Late Cretaceous volcanisms took place confined in Southern Korea, particularly actively in the southern part.

The Sindong Group yields in its lower part a guide taxon of the charophyta, *Clypeator jiuquanensis* [55], which has a known range of Hauterivian to lower Barremian [56]. It is, therefore, probable that the Sindong Group began to accumulate in the Hauterivian. The middle part of the Sindong Group yields the well-known *Trigonioides-Plicatounio* assemblage that allows correlation with strata containing the same biozone in Japan and China [57, 58, 59].

For the Hayang Group, an U-Pb CHIME isochron age, about 113.6 Ma, has been obtained from the zircon grains of the Kusandong Tuff in the top Haman Formation of the Hayang Group [60, 61]. It now appears that the boundary between the Haman and the Jindong Formations corresponds approximately to the Aptian-Albian transition (112 Ma). The Jindong Formation is the last, lacustrine phase of the Hayang sedimentation with dark-gray silty facies about 1500 m or more thick. The upper Aptian age of the Haman Formation and the lower Albian age of the Jindong Formation are based on this date.

The andesitic lower part (Jusasan Intermediate-Volcanic Subgroup) of the Yuchon Group shows an upper Albian-Cenomanian peak of the published isotope ages. Another, more distinct, peak of the published isotope ages, the Campanian-lower Maastrichtian peak appears to indicate the culmination of the Unmunsa volcanism [48, 49]. The top of the Yuchon Group may go up till the Paleocene in some sections according to isotope datings [18], but the post-Cretaceous part (volcanic rocks) is very minor in quantity. The Yuchon volcanics (except the basal part) and the comagmatic plutonic rocks are distinctly calc-alkaline, suggesting its subduction-related origin [17, 62].

It is well established that the Jasong and Kyongsang Synthems of Korea are entirely of continental deposits. The Jasong Synthem, in both Northern and Southern Korea, yield only fresh-water fossils, such as mollusks, estheriids and plant fossils [41, 51, 52].

The Kyongsang Supergroup have yielded only terrestrial fossils, plant leaves and stems, fresh-water mollusks, estheriid conchostracans, insects, tortoise, dinosaurs (both bones and footprints), bird tracks and charophytes. Channel beds, often with "shale clasts", poor sorting, mica flakes and the prevalent red rock color of the clastic rocks are additional evidence for the continental origin of the Kyongsang Supergroup [10, 11, 31, 32, 63]. Occurrences of nodular caliche suggest alternated wet and dry seasons over floodplains. Marls, usually less than 1 m thick, reflect swamp and temporary lake facies; thick black shale sequences yield lacustrine-paludal fossils. These environmental factors are also inherent in the Kyongsang Synthem in general.

In the Kyongsang Basin, the upper Albian Jindong Formation consists largely of black shales, admittedly a fresh-water anoxia event. The apparently coeval top part of the Hanbongsan Group (of the Daedonggang-Jaeryonggang Basin, North Korea) comprises a substantial reducing facies (Figure 2). Thus, the Jindong Formation and the uppermost part of the Hanbongsan Formation appear correlated to each other. On the other hand, it is certain that local tectonism significantly determines the sedimentary cycles, which differ even in parts of one and the same sedimentary basin; the Sindong and Hayang Groups show one major cycle in the northern part of the basin, while, in the southern part, it shows two cycles. The continental sedimentary cycles are determined by regional tectonic-climatic factor strongly affected by local vertical tectonics.

The Hauterivian-lower Albian sequence (Sindong and Hayang Groups) in the Kyongsang Basin is about twice thick and contains more conglomerates than in Northern Korean, showing a relative crustal unrest in Southern Korea. As to the same interval, no indigenous volcanic component has been reported in Northern Korea, but Aptian-lower Albian basalt and andesite notably alkaline occur in the Hayang Group and some acidic tuffs are very rarely intercalated in the Hauterivian-Barremian Sindong Group.

In the Inner Zone of Southwest Japan, the Tithonian-Valanginian sedimentary sequences in westernmost Honshu (Toyonishi Group) and in central Honshu (Itoshiro Subgroup of the Tetori Group) show no volcanism, only regressive facies from the shallow-marine or brackish-water below to fluvio-lacustrine above [64, 65].

The Kanmon Group of the Inner Zone of Southwest Japan and the Kyongsang Supergroup of the Kyongsang Basin of Southeast Korea show close affinity in fossils, age and sedimentary environment, and both basins were nearly adjacent prior to the opening of the East Sea (Sea of Japan). The Wakino Subgroup of the Kwanmon Group, consisted of four lithostratigraphic cycles each with basal conglomerates, is correlated with the Sindong Group on the basis of a common *Trigonioides-Plicatounio* Fauna [58]. An attempt was made to correlate each of these cycles with lithostratigraphic units of the Sindong Group, but such a correlation is not justified because the tectonics that control the cycles may be local rather than regional.

The U-Pb age of the middle-upper part of the Hayang Group is 113.6 Ma, and the K/Ar hornblende age of the vlocanic rocks of the Shimonoseki Subgroup is 105 and 107 Ma. These isotope ages support the correlation of the Hayang Group and the Shimonoseki Subgroup [66]. The Aptian-early Albian age of the Shimonoseki Subgroup is generally accepted by authors [67].

5. KOGURYO MAGAMTIC PROVINCE

In the Korean Peninsula, three Upper Mesozoic unconformity-bounded units are distributed with a distinct tendency that the lower unit occurs almost exclusively in Northern Korea, the middle unit occurs dominantly in Southern Korea and the upper unit occurs almost exclusively in Southern Korea, particularly in the southern part. The volcanic rocks in them also show a similar southeastward younging; they occur only in Northern Korea in the early epoch, while they occur only in Southern Korea in the last epoch. In general, the Late Mesozoic sedimentation and volcanism in the Korean Peninsula have shown a progressive younging toward the Pacific.

Such a tendency is also shown over the Koguryo Magmatic Province [8] recognized by a review of geologic maps of East Asia [1, 2, 3, 4] and various sources. Subcircular shaped, it covers Northeast China, Korea and parts of Russia and Japan (Figure 3), and is characterized by concentrations of the Jurassic-Cretaceous igneous rocks and sedimentary basins proving intimate genetic relations of sedimentary basins and magmatisms. Ancient nation Koguryo was in about the center of this province.

The Koguryo Magmatic Province was already extant in the Jurassic Period when post-Indosinian S-type granites of a calk-alkaline composition [4, 18] emplaced there extensively. The province is a part of the Jurassic-Cretaceous peri-Pacific magmatic belt of the Asian continental margin, but its subcircular shape mimics a hot spot above an enduring plume. It is notably located in the eastern part of the Tienshan-Mongolian belt and adjacent parts of the Sino-Korean and the Siberian continents. Magamtisms were continual in this province during the Jurassic and the Cretaceous and enven thereafter [68]. The persistent magmatisms may have been caused by a mantle plume below, but a subduction zone was also closely located. All the Jurassic-Cretaceous magmatic rocks except the Aptian-Albian ones are reportedly calc-alkaline [12, 14, 18] implying ample water supply for the magma genesis. The subduction of oceanic plates may have been a source of water, but the Tienshan-Mongolian belt must have had incipient water derived from the Paleozoic ocean that formed the Tienshan-Mongolian belt. It is possible that mantle plume and the nearby subductions have mutually interfered to produce this magmatic province.

Notable is the location of the province in the crossing juncture of the peri-Pacific Asian continental margin and the Tienshan-Mongolian belt. Such a geologic location of the province allows an interpretation that the magma genesis and ascention was facilitated by the heterogeneous, loose and water-lain aggregation of the Tienshan-Mongolian belt and some fracture zones developed in adjacent cratons.

In the eastern part of the Tienshan-Mongolian belt, four micro-continental blocks are recognized: Buryea, Khanka, Songliao and Jiamusi Blocks. During the Cretaceous Period the Buryea and Khanka Blocks were exposed but the Songliao and Jiamusi Blocks were sites of sedimentation [3]. The Late Jurassic-Early Cretaceous volcanisms had a tendency of avoiding these micro-continental blocks.

Among Upper Jurassic-lowermost Cretaceous andesitic-basaltic volcanic belts (Figure 3) two of them lie in the Tienshan-Mongolian belt saliently between continental blocks. Another one located in the northern margin of the Sino-Korean continent is here called the North Sino-Korean Volcanic Belt (NSKVB), which raises a particular concern because the North Korean Volcanic Belt (NKVB) is adjacent to this belt. All these volcanic belts are discordant with the Pacific margin. Notable is the NSKVB lying perpendicular, not parallel at all, to the Pacific

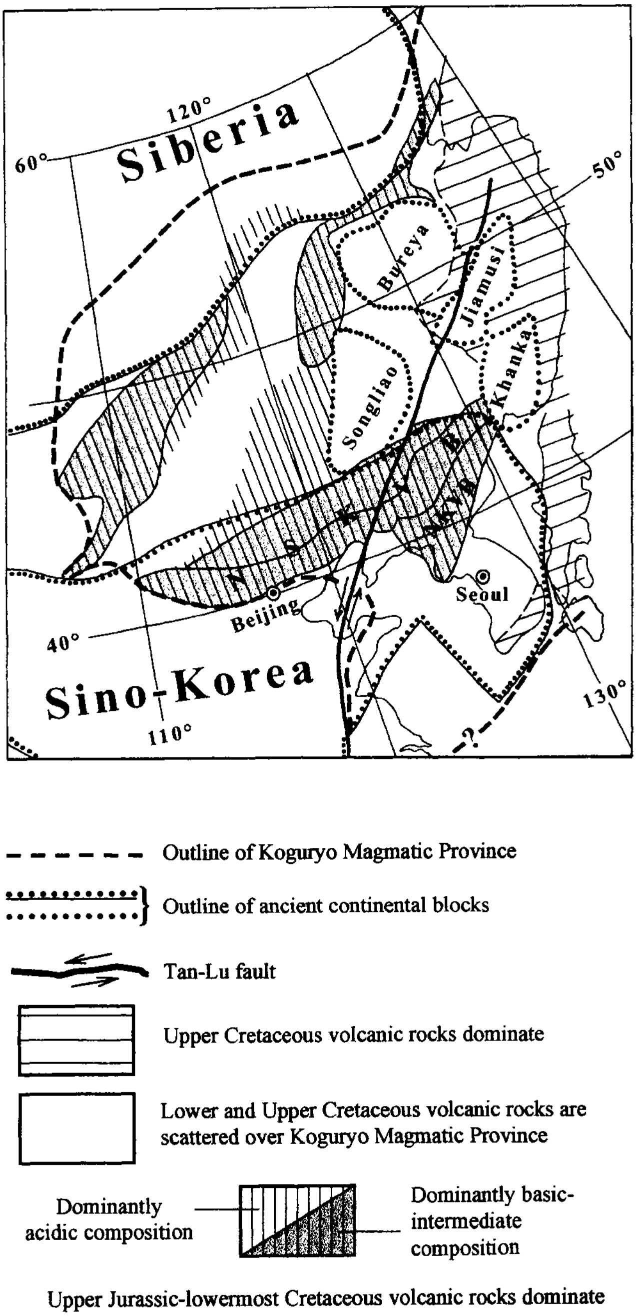

Figure 3. Koguryo Magamtic Province with concentrated Jurassic and Cretaceous magmatisms. Four major ancient massifs or micro-continents in the eastern part of the Tienshan-Mongolian Zone are also shown, though only Bureya and Khanka were exposed during the Cretaceous. Cretaceous volcanisms took place unevenly over the province. The Upper Jurassic-lower-most Cretaceous volcanic belts are emphasized on the map to show the location of the North Korean Volcanic Belt (NKVB), which is adjacent to the North Sino-Korean Volcanic Belt (NSKVB). Volcanisms (andesites and basalts) were particularly active on the southern belt of NSKVB suggesting eruptions through fractures developed along the margin of the Sino-Korea Craton.

margin. Its location suggests a reactivation of an ancient Sino-Korean continental margin; a new-born fracture system appears served as the outlet of the volcanism. The NSKVB may be divided into two zones as shown in Figure 3. The southern zone is the most active volcanic chain probably corresponding to the fracture system that developed in parallel with the suture between the Tienshan-Mongolian belt and the Sino-Korean continent. The NKVB is relatively sparse with vocanic rocks.

During the Late Jurassic-Early Cretaceous time, the volcanics of acidic compositions also erupted. All these coeval basic to acidic volcanic rocks occur largely in the western part of the Koguryo Magmatic Province (Figure 3). In the succeeding Cretaceous time, volcanisms of various compositions took place unevenly in the Koguryo Magmatic Province, but their general locations show an eastward migration of magmatism compared to the earlier volcanisms. Though the Late Cretaceous volcanisms took place in various parts of the Koguryo Magmatic Province, they were particularly active along the linear eastern periphery of the Asian continental margin, strongly suggesting, together with its calc-alcaline composition, a subduction related origin. It thus appears that the Late Jurassic-Cretaceous volcanisms within the Koguryo Magmatic Province generally shifted with time toward the east, probably in response to the progressive retreat of the subduction zone toward the Pacific.

Why, on the other hand, magmatisms were sustained in the same overall area throughout the Jurassic and thereafter [68] ? The concept of a plume may be the answer. But, since the plume-like entity can also possibly be caused by a subduction system, a further answer should await special studies.

6. CONCLUDING REMARKS

The Jasong Synthem occurs almost exclusively in Northern Korea, but the Valanginian disturbance affected a migration of the depocenter toward the Kyongsang Basin in Southeast Korea. Throughout the Late Mesozoic, the crustal unrest shown by sedimentation and volcanism generally propagated from north to southeast; the sedimentary volume decreased with time in Northern Korea, while it increased in Southern Korea; and, volcanism was confined in the lower horizons (Upper Jurassic-lowermost Cretaceous) in Northern Korea, while they increased with time in Southern Korea.

The North Korean volcanic belt in the time of the Jasong Synthem was adjacent to a more extensive and intensive coeval volcanic belt (the southern belt of NSKVB in Figure 3) that developed in the northern margin of the Paleozoic Sino-Korean continent. The location of the sublatitudinal Northern Sino-Korean Volcanic Belt (NSKVB) suggests its eruption through fractures reactivatedly developed in parallel with the suture between the Tienshan-Mongolian belt and the Sino-Korean continent. The trend of this volcanic belt, perpendicular to the Pacific margin, is alien from a subduction-caused genesis. The Koguryo Magmatic Province was already present in the Jurassic Period when post-Indosinian S-type granites of a calc-alkaline composition [4, 18] emplaced there extensively. The long life and the shape of the province suggest its plume-generated origin. But, on the other hand, the magmatisms show a vague eastward younging implying a relation with subduction processes. Therefore, both a mantle plume and also the subductions of the oceanic plates may have mutually contributed to create the magmatic zone. The intimate association in space and time of the volcanic belts and sedimentary basins reflects their genetic relation. A crustal mobility would be manifested as magmatisms and sedimentations, and it is likely that the magma genesis and migration have a critical role in the origin of sedimentary basins.

The Late Cretaceous volcanism along the very margin of Asia, which is dominantly I-type and calc-alkaline [4], is explained to have been caused solely by the subduction of the oceanic plate. The upper Albian-Upper Cretaceous basins and the basin-fills, that are dominated by volcanic rocks, occur most commonly in the southern part of South Korea in a volcano-sedimentary belt running WSW-ENE parallel to the southern coastline of the Korean Peninsula. The mid- to Upper Cretaceous volcanic rocks were most intensely erupted and emplaced in this volcanic belt about 150 km wide.

The Kyongsang Synthem of whole Korea has now been subdivided into two subsynthems. The unconformity between the clastics-dominant Lower Kyongsang Subsynthem (Aptian-lower Albian) and the volcanics-dominant Upper Kyongsang Subsynthem (upper Albian-Paleocene) appears interregionally significant. It marks a new geological regime characterized by a series of volcanisms caused by a new subduction system.

According to Engebretson et al. [25], the Izanagi plate of the Pacific Basin was spreading toward the north almost parallel with the East Asian continental margin in the Aptian-Albian time. Since a margin-parallel spreading means no subductions, a compressional tectonics is not expected. The minor, alkaline volcanism of the Aptian-Albian interval was succeeded by a major, calc-alkaline one in the Upper Cretaceous in Korea. If the former volcanisms were caused by intra-plate volcanisms during the cessation of the subduction, the latter volcanisms may have been caused by an intensive volcanism caused by speeded subduction when the subduction was restored after a cessation of the subduction [24]. The pulling effect of the coast-parallel spreading of the oceanic plate toward the northeast can explain the sinistral slicing of the South Korean crust and the resultant Aptian-Albian pull-apart basins with some alkaline volcanisms.

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The Cretaceous System of the Japanese Islands and its physical environments

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The lithologic features of the Cretaceous sequence in the Japanese Islands are summarized, according to the most recent data, and its depositional environments are discussed in the context of tectonics and sedimentology. This paper focusses on the tectonic change of the East Asian continental margin from the passive margin in the latest Jurassic and the transformed margin in the late Early Cretaceous, to an oblique-slip convergent margin in the Late Cretaceous. This change furnished variety of sedimentary basins and igneous activity with many characteristic features.

1. INTRODUCTION

The general outline of the Cretaceous System of the Japanese Islands is provided by Matsumoto [1, 2, 3], Tanaka [4] and Hayami and Yoshida [5]. In addition, international correlation and lithologic features of the Japanese Cretaceous have also been summarized in Matsumoto [6], Hirano [7], Matsukawa et al. [8] and Okada et al. [9]. However, in the past few years abundant new data in the fields of stratigraphy, sedimentology, paleontology, petrology and chronology of the Cretaceous rocks have been added.

In this article, the general features of the Japanese Cretaceous are briefly summarized, based on recent information, and their environmental significance is discussed.

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2. DISTRIBUTION OF CRETACEOUS STRATA AND GEOLOGIC SETTING

The Japanese Islands are divided geologically into two major regions by the Itoigawa-Shizuoka Tectonic Line: Northeast and Southwest Japan. Northeast Japan is further subdivided by the the Tanakura Tectonic Line in Northeast Honshu and by some tectonic lines in Hokkaido. Southwest Japan is subdivided into the Inner Zone and the Outer Zone by the Median Tectonic Line. It is agreed that both the Tanakura and Median Tectonic Lines came into existence during the Cretaceous.

The distribution of the Cretaceous rocks conforms with these tectonic frameworks, as shown in Figures 1, 2 and 3. The stratigraphy and correlation of major Cretaceous units in the Japanese Islands are given in Figure 4.

2.1. Northeast Japan

2.1.1. Hokkaido

Hokkaido Island is tectonically divided into seven major belts [10] (Figure 1) and the distribution of Cretaceous rocks and their lithologic features are summarized as follows:

2.1.1.1. Nemuro Belt

The Nemuro Belt occupies the easternmost part of Hokkaido and is mainly represented by the Nemuro Group (Campanian to Eocene) (Figures 1, 4), which is composed of andesitic and basaltic volcanic conglomerates and breccias, volcanic sandstones and mudstones [11, 12] with intercalations of alkaline basaltic sills with pillowed structure [13]. The K/T boundary lies within the succession of the Nemuro Group (Figure 4).

2.1.1.2. Tokoro Belt

The Tokoro Belt consists of three units: the Upper Jurassic to Upper Cretaceous Nikoro Group and the Campanian to Paleocene Saroma and Yubetsu Groups (Figure 4). The Nikoro Group, more than 4000 m thick, is characterized by an ophiolite sequence of basaltic pillowed lava flows, basic hyaloclastites and pyroclastics accompanied by radiolarian chert and micritic limestone[14]. The Saroma Group, about 1500 m thick, overlying the Nikoro Group unconformably, consists of terrigenous turbidites. The Yubetsu Group, more than 10000 m thick, which is almost contemporaneous with the Saroma Group, is composed of terrigenous and volcanic turbidites [15].

2.1.1.3. Hidaka Belt

The Hidaka Belt is subdivisible into the Axial Subbelt and the Idonnappu Subbelt in the western marginal part [16] (Figure 1). The Axial Subbelt is represented by the Valanginian to Paleocene Hidaka Supergroup in the central and northern parts of the belt and the Campanian to Paleocene Nakanogawa Group in the eastern part. The Hidaka Supergroup comprises the complex of turbidite sandstone and mudstone associated with melanges of chert, limestone and green rocks. The Nakanogawa Group, about 9000 m thick, consists of terrigenous turbidites with melanges [17, 18]. Olistoliths of these melanges contain Permian and Triassic fossils.

The Idonnappu Subbelt is represented by melange facies, which consists of blocks of chert,

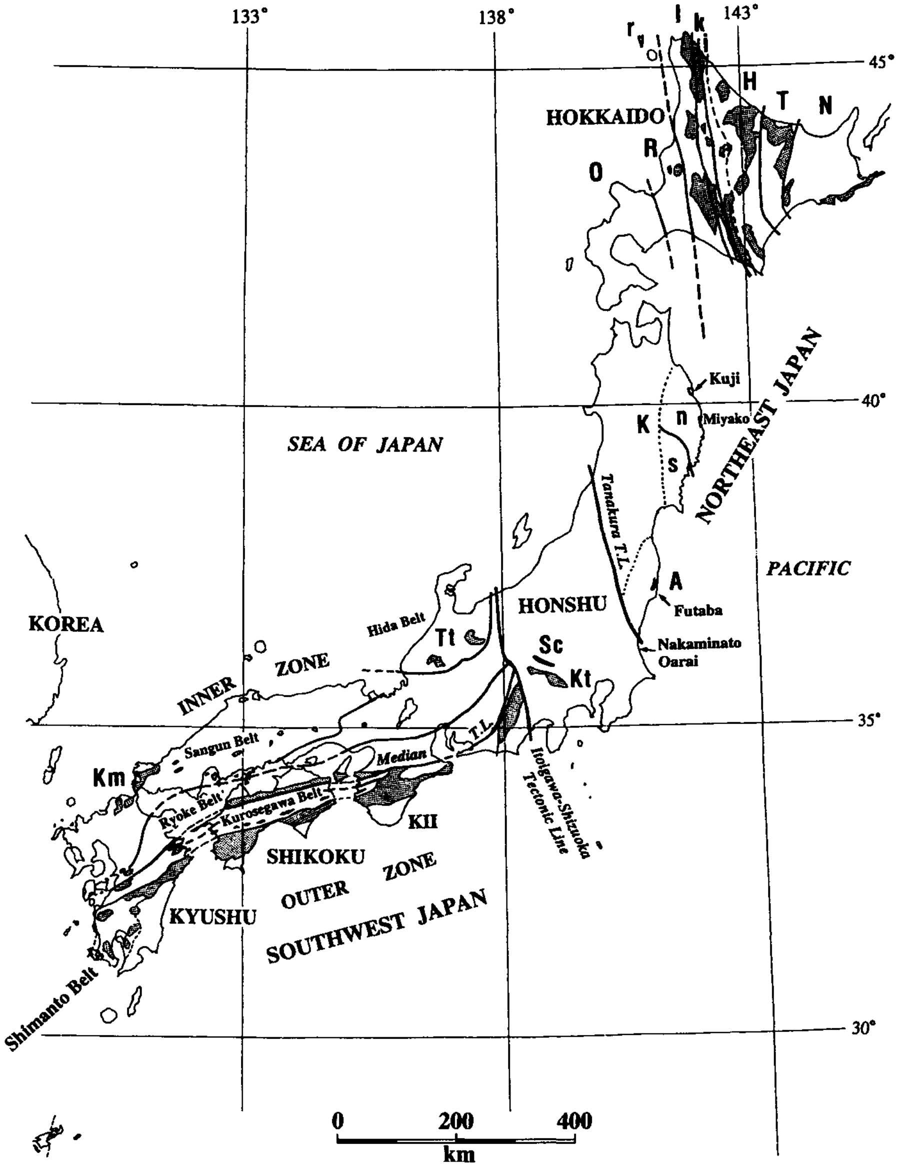

Figure 1. Major tectonic divisions of the Japanese Islands showing the distribution of Cretaceous strata. For Hokkaido, N: Nemuro Belt, T: Tokoro Belt, H: Hidaka Belt, i: Idonnappu Subbelt, I: Ishikari Belt, R: Rebun-Kabato Belt, O: Oshima Belt, r: Rebun Island. For Northeast Japan, K: Kitakami Massif, n: North Kitakami Belt, s: South Kitakami Belt, A: Abukuma Massif, Sc: Sanchu Graben, Kt: Kanto Mountains. For Southwest Japan, Tt: Tetori, Km: Kanmon.

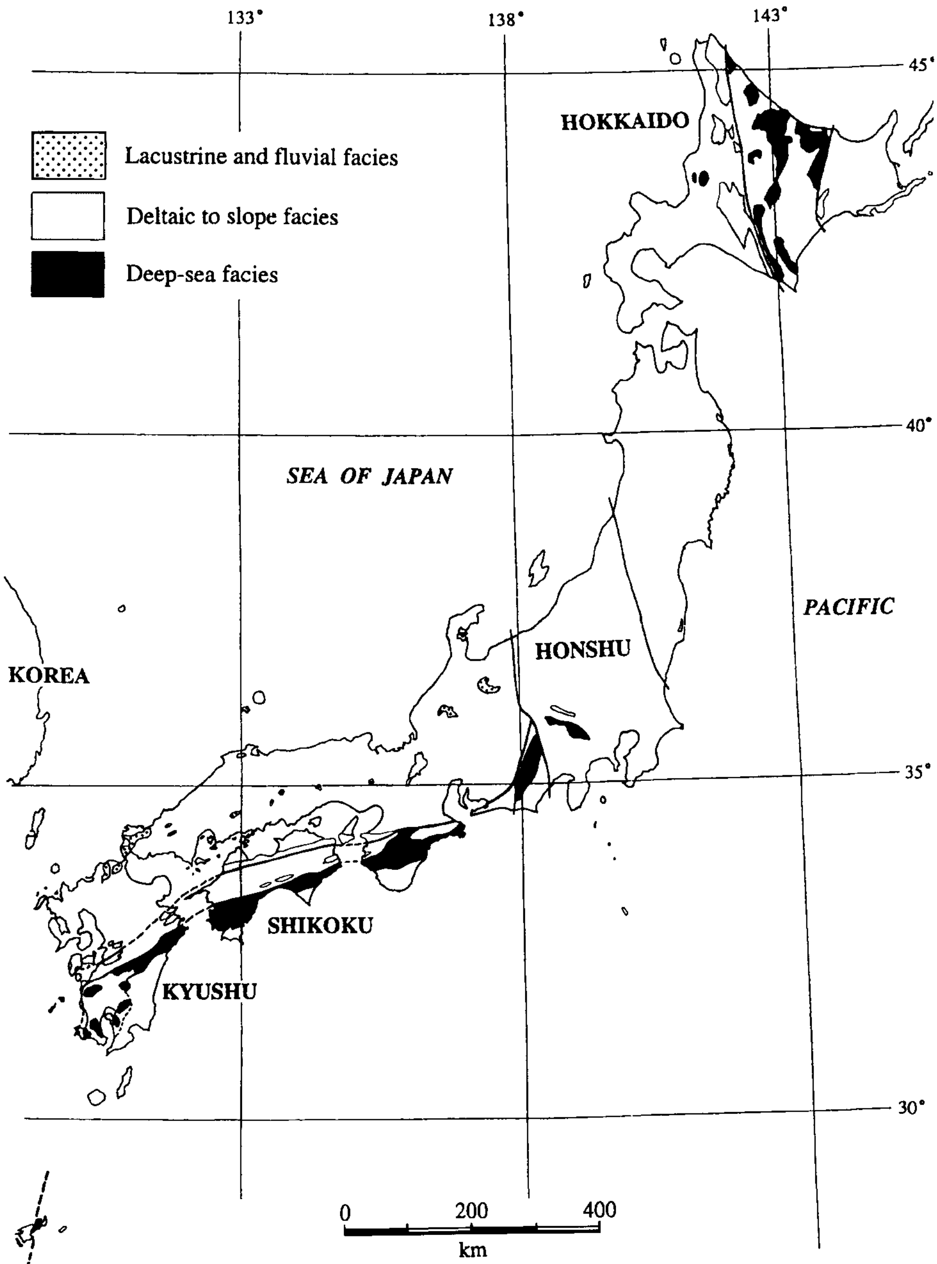

Figure 2. Discrimination of Lower Cretaceous sedimentary facies in the Japanese Islands.

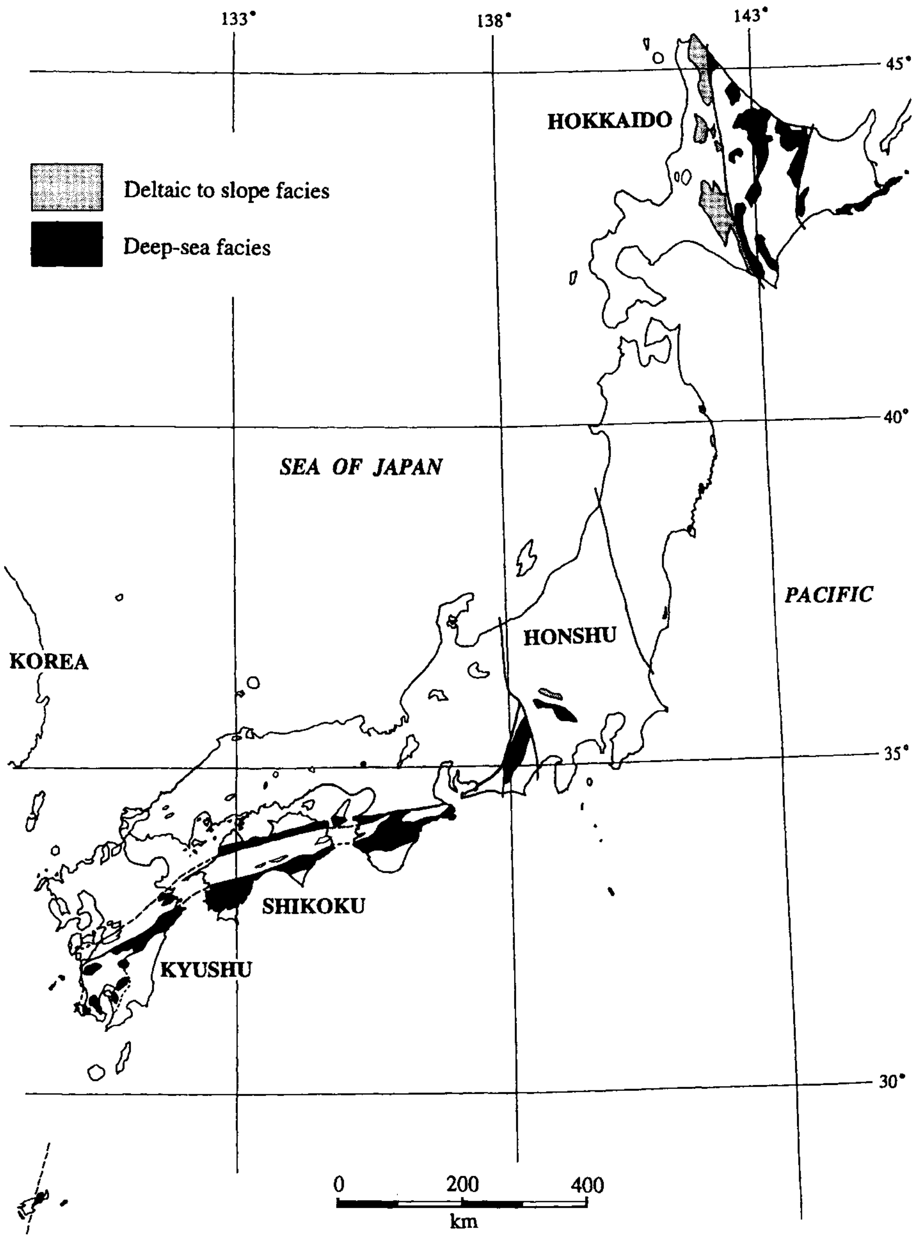

Figure 3. Discrimination of Upper Cretaceous sedimentary facies in the Japanese Islands.

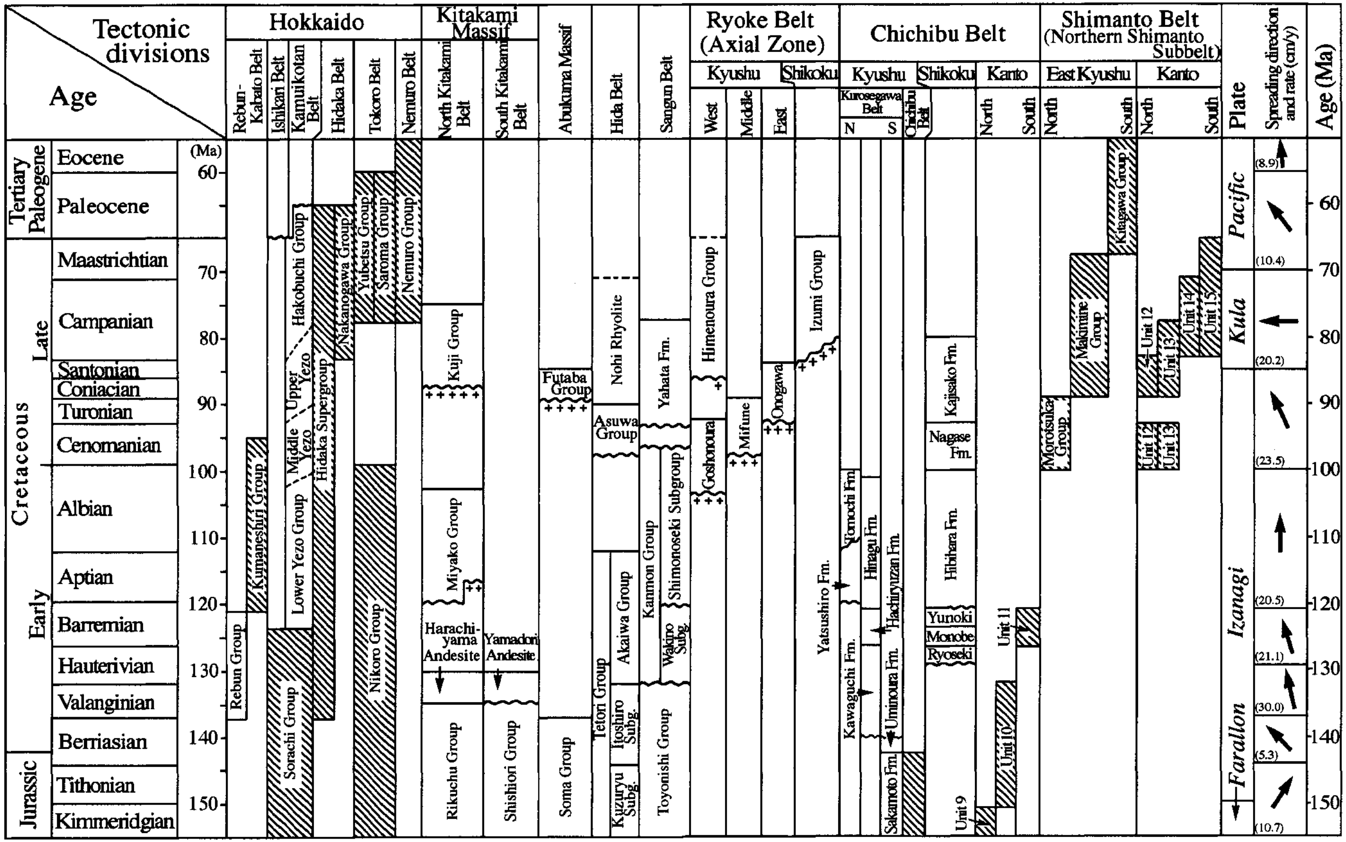

Figure 4. Stratigraphy of major Cretaceous units in the Japanese Islands and their correlation. For Units 9 - 15 in Kanto, refer to Takahashi and Ishii [81] and Takahashi [108]. Informations of plate arrangements are from Seno and Maruyama [82].

limestone and green rocks embedded in the Hauterivian to Campanian matrix [16].

2.1.1.4. Kamuikotan Belt

This belt is characterized by the Kamuikotan high P/T metamorphic rocks and ultrabasic rocks, which are accompanied by the ophiolitic Sorachi Group (late Tithonian to Albian) (Figure 4). The Kamuikotan metamorphic rocks show the ages of three groups, 108-145, 91-107 and 50-84 Ma [19].

2.1.1.5. Ishikari Belt

The Ishikari Belt is mainly represented by the Yezo Supergroup from Aptian to Maastrichtian in age (Figure 4). It is subdivided into the Lower Yezo, Middle Yezo, Upper Yezo and Hakobuchi Groups [10, 20].

The Lower Yezo Group, 1000-1300 m thick, late Barremian to late Albian, is characterized by turbidite facies, in which dislocated blocks [10] of shallow-water carbonates with corals, rudists, an oyster, gastropods, calcareous algae and an orbitolinid foraminifer [21] are sporadically found. The Middle Yezo Group, 1800-2000 m thick, late Albian to Turonian, is characterized by shallow-marine coarse-grained clastic facies in the western part and offshore muddy facies in the eastern part of the belt. A detailed sequence-stratigraphic analysis was applied by Ando [22, 23] to the Mikasa Formation of the sandstone-dominated upper sequence of the Middle Yezo Group. Furthermore, at the Cenomanian-Turonian boundary in the MiddleYezo Group, the black shale bed indicating the oceanic reduction conditions has been detected by Hirano et al. [24].

The Upper Yezo Group, 400-1300 m thick, Coniacian to Santonian, shows argillaceous facies and marks the highest transgression called the Urakawan Transgression [25, 26].

The Hakobuchi Group, 400-900 m thick, Campanian to Maastrichtian, consists of coarse-grained, thick-bedded sandstones with shallow-sea molluscs, sometimes intercalated with thin coal seams [27].

2.1.1.6. Rebun-Kabato Belt

This belt includes important Cretaceous outcrops of Rebun Island and the Kabato Mountains (Figure 1). Rebun Island is represented by the Rebun Group comprising Valanginian to Barremian andesitic volcanics with ammonite-bearing marine clastic layers and limestone lenses [28]. The Kabato Mountains are constituted by the Lower Cretaceous Kumaneshiri Group, which is composed of radiolaria-bearing siliceous shale, sandstone, and andesitic to basaltic pyroclastics [29] (Figure 4).

2.1.1.7. Oshima Belt

This belt (Figure 1) is characterized by Jurassic to Early Cretaceous accretionary prisms, which contain olistoliths of Late Carboniferous to Triassic black shale, siliceous shale, chert and limestone with radiolarians and conodonts [30, 31]. In this belt granites are of the Cretaceous components.

2.1.2. Geologic setting of Hokkaido

During the Cretaceous, Hokkaido Island was separated into two tectonic masses: the Yezo Arc-Trench System to the west and the Okhotsk Block to the east, between which the Sorachi Ocean, a part of the Izanagi-Kula Plate, was present [10]. The general tectonic setting of Hokkaido is summarized below.

In the Early Cretaceous time, the Rebun Belt constituted a magmatic front along the eastern margin of the Eurasian continent, which extended south to the Kitakami Massif [32]. As a forearc basin of this magmatic arc, the Yezo Supergroup basin in Hokkaido and some small coastal basins in the Kitakami Massif were developed.

The Yezo Supergroup was deposited mainly in a forearc basin and in the inner trench slope to trench environments. Thus, coarse-grained paralic to shallow-sea sediments were deposited at the western margin of the forearc basin, while the muddy sequence with frequent turbidites was formed in the trench-slope to trench areas. The Cenomanian/Turonian oceanic anoxic event (C/T-OAE) is recorded in the Yezo forearc basin [24, 33].

"*Orbitolina* limestone" blocks sporadically found in the turbidite facies of the Lower Yezo Group may have been dislocated as olistoliths either from the western shelf margins or from nearby tectonic highs or from the trench-slope break, where "*Orbitolina* limestone" was deposited initially as a part of reefal limestones in the tropical-subtropical realm [21]. Beneath these cap rocks, the tectonic highs were underseated by the Kamuikotan metamorphic rocks, which were generated from ophiolites and melanges due to the northwestward subduction of the Izanagi Plate in 100-125 Ma. Serpentinite masses associated with these metamorphic rocks seem to have been uplifted in the same diapiric process as in the modern Mariana forearc [34].

Beyond the Yezo Trench, ocean-floor pelagic sediments and submarine volcanic rocks (Sorachi Group) constituted the floor of the Sorachi Ocean. According to microfossil biostratigraphic and paleomagnetic data of the trapped Izanagi oceanic rocks, limestone and chert blocks from the Tokoro Belt in eastern Hokkaido show a mean inclination of $I=3.4^\circ$, corresponding to a paleolatitude ranging from 5° N to 9° S for Callovian to Aptian times (168-112 Ma) [35]. This implies that the Tokoro oceanic rocks were formed in a seamount environment near a spreading ridge or a hotspot in the paleoequator region on the Izanagi Plate. Later northward migration of these rocks to the present position is well explained by the models of Engebretson et al. [36] and Henderson et al. [37].

On the opposite side of the Sorachi Ocean, there was a landmass on the Okhotsk Plate in the Late Cretaceous time [10, 38], and east-dipping subduction started on the west side of the paleoland. This means that the trapped Kula Plate began rifting both eastwards and westwards. Thus, ophiolitic sequence originated at low latitudes were accreted to the Okhotsk Paleoland, forming an arc-trench system with eastward subduction. This tectonic situation gave rise to highly volcanic sediments in forearc and trench areas [39].

Further in the latest Cretaceous, intense subduction took place along the southern margin of the Okhotsk Paleoland, producing the proto-Kuril Trench. This subduction triggered an intense volcanism which produced the volcanogenic Nemuro Group.

The proto-Kuril Arc in the Okhotsk Plate and the Yezo Arc-Trench System collided with each other in the late Miocene [32, 38].

2.2. Northeast Honshu

Cretaceous rocks in Northeast Honshu are mainly exposed in the Kitakami and Abukuma Massifs (Figure 1).

2.2.1. Kitakami Massif

This tectonic block is subdivisible into the North Kitakami Belt and South Kitakami Belt (Figure 1). In the North Kitakami Belt the uppermost Jurassic to Hauterivian sediments called the Rikuchu Group (about 2500 m thick), the Aptian to Albian Miyako Group (about 200 m thick) and the Santonian to Campanian Kuji Group (400-600 m thick) are exposed in a narrow belt along the Pacific coast.

The Rikuchu Group is mainly characterized by dacitic-andesitic rocks associated with sandstone and shale [40]. This sequence is intruded by granites of 121 ± 6 Ma and 128 ± 17 Ma [41] and is covered by the Harachiyama Andesite.

The Miyako Group is charactrized by littoral to neritic deposits of conglomerates and calcareous sandstones, which contain varieties of the Urgonian fauna [42]. Sano [43] discovered a biohermal limestone with a coral-rudist framework in the Miyako Group, suggesting a high-energy environment of deposition on a narrow rimmed shelf in tropical to subtropical conditions [21]. The Kuji Group lying on Early Cretaceous granites is dominated by cross-bedded sandstones and accompanied by conglomerate, coaly shale and felsic tuff [44]. Ammonites and inocerami are found in the middle sequence and plant remains in the upper.

In the South Kitakami Belt Lower Cretaceous sequence of shallow-marine to paralic facies with some nonmarine beds is sporadically exposed.

2.2.2. Abukuma Massif

Along the Pacific coast narrow exposures of the Berriasian to Valanginian Soma Group (about 180 m thick) and the Coniacian to Santonian Futaba Group (about 350 m thick) are distributed. The Futaba Group overlies Early Cretaceous granites and is composed of fluvial to shallow-marine deposits [45].

2.2.3. Kanto Block

As the Kanto Block located between the Tanakura Tectonic Line and the Itoigawa-Shizuoka Tectonic Line shows a geologic continuity from Southwest Japan (Figure 1), the Cretaceous System exposed in this block is described in the Outer Zone of Southwest Japan.

2.2.4. Geologic setting of Northeast Japan

The North Kitakami Terrane is composed of a Jurassic subduction complex that collided with the South Kitakami Terrane in the earliest Cretaceous time (Berriasian to Valanginian) [46]. The South Kitakami Terrane was originally a part of a Gondwanaland-related continent [46, 47]. Since that collision, the Kitakami Massif has been subducted westwards by the proto-Pacific plates [46]. This subduction gave rise to intense andesitic volcanism along the eastern margin of the Kitakami Massif, which was connected with the Rebun-Kabato Belt in Hokkaido in the Early Crtaceous time, as already mentioned.

This collision hypothesis has recently been criticized by Tazawa [48, 49] that Triassic pelecypod faunas from the South Kitakami Terrane are characteristic of the Boreal type and are at variance with a view of lower latitude origin of the fauna.

2.3. Inner Zone of Southwest Japan and its geologic setting

The Inner Zone of Southwest Japan is tectonically divisible into the Hida, Sangun and Ryoke Belts (Figure 1). The most important feature of the Inner Zone of Southwest Japan is the continental geologic characteristics such as a prevalence of lacustrine to fluvial red beds, quartz-arenite sediments and occurrence of plenty of plant fossils and fresh-water fauna, and rather simple structural deformations of strata, as described below:

2.3.1. Hida Belt

This belt is represented by the Lower Cretaceous Tetori Group and the Upper Cretaceous Nohi Rhyolite (Figure 4). The Cretaceous sequence of the former is subdivided into the Itoshiro Subgroup (300-1300 m thick) in the lower and the Akaiwa Subgroup (600-1600 m thick) in the upper [50]. The Itoshiro Subgroup (Berriasian to Valanginian) consists of conglomerate, feldspathic-quartzose sandstone and mudstone of brackish-water to freshwater facies with abundant bivalve assemblages such as *Myrene tetoriensis, Tetoria yokoyamai, Sphaerium* sp. and so on [51, 52, 53]. It also yields plant fossils known as the Tetori Flora [54, 55].

The Akaiwa Subgroup (Hauterivian - Aptian) is composed of conglomerate, feldspathic-quartzose sandstone and mudstone of lacustrine origin, containing abundant fossils of *Plicatounio naktongensis, Trigonioides tetoriensis, Nagdongia soni,* and so on [51, 52, 53], together with plant fossils of the Tetori Flora [54, 55].

The Tetori Group is characterized by fresh-water to brackish-water and partly marine lithofacies under fluvial, lacustrine and deltaic environments [52]. Fossils of molluscs (Tetori Fauna) [51, 53], terrestrial plant remains (Tetori Flora) [55], and dinosaur bones [56, 57] and footprints [8, 57, 58] indicate the same paleobiogeographical province of this area as Northeast China and South Korea during the Early Cretaceous time [58].

Matsukawa and Obata [57] and Matsukawa et al. [52, 53] delineated the paleogeography of the Tetori Basin that it opened to the north facing the proto-Pacific in Late Jurassic to early Early Cretaceous under the influence of cold currents and the Tetori embayment became more pronounced in Barremian time.

The ‘orthoquartzite" clasts commonly found in the Tetori Group give the K-Ar age of 461-783 Ma [59]. This means that these pebbles are older than the oldest stratigraphic units in Japan, suggesting their continental origin [60].

The Nohi Rhyolite (Coniacian to Campanian) is extremely voluminous and mainly comprises welded tuff of rhyolitic to dacitic composition [61].

2.3.2. Sangun Belt

The Sangun Belt occupies the west part of Honshu and the northwestern part of Kyushu (Figure 1). The westernmost region of Honshu is characterized by extensive distribution of Jurassic to Cretaceous sediments and pyroclastic rocks. Among them most important are the Late Jurassic to Valanginian Toyonishi Group (600-900 m thick) and the Hauterivian to

Cenomanian Kanmon Group (3500 m thick) [62] (Figure 4).

The Toyonishi Group is divided into the Kiyosue Formation (450-550 m thick) in the lower and the Yoshimo Formation (200-300 m thick) in the upper part [63]. The former is rich in plant fossils of the Kiyosue Flora and the latter is characterized by the occurrence of many beds of orthoquartzite [64], yielding abundant brackish-water molluscs, which are common with the Ryoseki Fauna in the Outer Zone of Southwest Japan [5, 65]. The plant fossils from this group are defined as the Mixed-type Flora between the Tetori-type and the Ryoseki-type Floras [55].

In particular, brackish-water to paralic sediments composed of quartz arenite or orthoquartzite suggest strongly that these quartzose sediments were deposited in close relation to the Korean continent, where Cambrian to Ordovician quartz arenites are common [66].

The Kanmon Group overlies the Toyonishi Group disconformably, and is the most important geologic unit in the Inner Zone. The Kanmon Group is characterized by varicolored sediments, being subdivided into the lower Wakino Subgroup and the upper Shimonoseki Subgroup (Figure 4).

The Wakino Subgroup, about 1200 m thick, consists of unsorted conglomerate, red-colored sandstone and mudstone of alluvio-fluvial environments and turbidite and rhythmite facies of the offshore lacustrine environment [67]. It is characterised by abundant nonmarine molluscs such as *Plicatounio naktongensis, P. kwanmonensis, Trigonioides intermedia, T. suzukii, Sphaerium coreanicum,* and others as pelecypods and *Brotiopsis wakinoensis, B. kobayashii* and *Viviparus* sp. as gastropods [68] in addition to estherids, algal stromatolites, fish [69], birds, and tortoise and dinosaur-bone fragments. This fresh-water molluscan fauna is quite similar to the Sindong Fauna of the Kyongsang Supergroup in South Korea[70].

The well-preserved freshwater-fish fauna from the Subgroup is charactrized by the abundance of osteoglossiformes and clupeiformes, suggesting many similarities to the fish fauna in south eastern China [69].

The Shimonoseki Subgroup, about 2300 m thick, is almost entirely volcanogenic (mostly andesitic and porphyritic), comprising lavas, pyroclastics and alluvial sediments. Volcanic breccias and conglomerates indicate a debris flow origin. Fossils are very rare, except for *Nippononaia* and *Estherites* species. The lithology of the Subgroup is comparable to that of the Yuchon Group of the uppermost Kyonsang Supergroup in South Korea, which is characterized by the dominance of andesite and rhyolite [70, 71].

The Kanmon Group is overlain unconformably by the Yahata Formation, which consists of dacitic to rhyolitic pyroclastics and andesitic lava flows. This formation is correlated in age to the Nohi Rhyolite in central Honshu.

The Cretaceous rocks in the Sangun Belt are closely related to those in South Korea.

2.3.3. Ryoke Belt

This belt is characterized by the basement of the Ryoke granitic rocks of the age of 140 - 70 Ma and the low P/T Ryoke metamorphic rocks of 120 - 50 Ma in age. In this belt Cretaceous sedimentary basins are well developed as half-graben depressions to the north of and along the Median Tectonic Line [72, 73] (Figures 1, 4, 5, 6). The term of the Axial Zone is often applied to the sedimentary basin areas developed in the southernmost part of the

Ryoke Belt along the Median Tectonic Line and its derivatives mainly in Kyushu and Shikoku Islands.

These basins accommodate relatively thick clastic wedges from nonmarine to shallow-sea facies as well as deep-sea facies. Deposition of the sediments started in the latest Albian to earliest Turonian in the far west (Goshonoura Group), in the Cenomanian to Turonian at the central part of Kyushu (Mifune Group), and in the Turonian to Santonian at the eastern part of Kyushu (Onogawa Group), and in the latest Campanian to Maastrichtian in northern Shikoku, Awaji Island and the Kii Peninsula, showing the gradual shift of depocenters and the younging trend to the east (Figure 5).

The Goshonoura and Mifune Groups are characterized by such unique molluscan fauna of shallow-sea to brackish-water environments as *Trigonioides (Kumamotoa) mifunensis, Matsumotoa japonica* and *Pseudasaphis* sp. They are also characterized by the presence of red beds and, in particular, the upper sequence of the Mifune Group consists of more than 1000 m thick nonmarine red beds, in which fragments of dinosaur bones are found.

The Onogawa and Izumi Groups show an enormous thickness, more than 20000 m for the Onogawa Group and 4000 to 7000 m for the Izumi Group, both of which are characterized by marine turbidite facies.

These clastic sediments reflect local geology around the basins as their sources [74]. These basins show a common tectonic feature of strike-slip origin.

In the Kanto Block farther to the east, isolated exposures of the equivalent formation of the Izumi Group called the Nakaminato Group is found (Figure 1), which consists also of marine turbidites more than 3000 m thick.

2.4. Outer Zone of Southwest Japan and its geologic setting

2.4.1. Kurosegawa Belt

The Kurosegawa Belt occupies the innermost part of the tripartite "Chichibu Belt" in the Outer Zone of Southwest Japan, extending from Kyushu Island through Shikoku Island to the Kanto Mountains west of Tokyo (Figure 1).

Cretaceous basins in this belt are narrowly and intermittently distributed. Sediments range in thickness from 300 to 3000 m, showing two types of sedimentary facies: shallow-sea to brackish-water facies and deep-sea turbidite facies.

Most of the strata in this belt show fossiliferous shallow-sea to brackish-water facies, which are represented by the Ryoseki Group (300-500 m thick, Valanginian to Barremian) in Shikoku and its equivalent Kawaguchi Formation (400 m thick) in West Kyushu, Tatsukawa, Hanoura, Hoji and Fujikawa Formations (800-1000 m thick altogether, Hauterivian to Albian) in East Shikoku [75], and the Sanchu Cretaceous (more than 2700 m thick, Hauterivian to Cenomanian) in the "Sanchu Graben" in the Kanto Mountains [76].

The Ryoseki sediments are characterized by quartzose-feldspathic sandstones and red beds in some parts as well as by the Ryoseki Flora [55] and the Ryoseki Fauna of brackish-water shells such as *Eomiodon, Costocyrena, Hayamina, Tetoria,* and *Protocardia* species [5].

A deep-sea turbidite facies is developed in some small basins in West Kyushu, in which the Berriasian to Hauterivian Uminoura Formation (150 m thick), the Hauterivian to Barremian Hachiryuzan Formation (300 m thick), Aptian to Albian Hinagu Formation (800 m), and

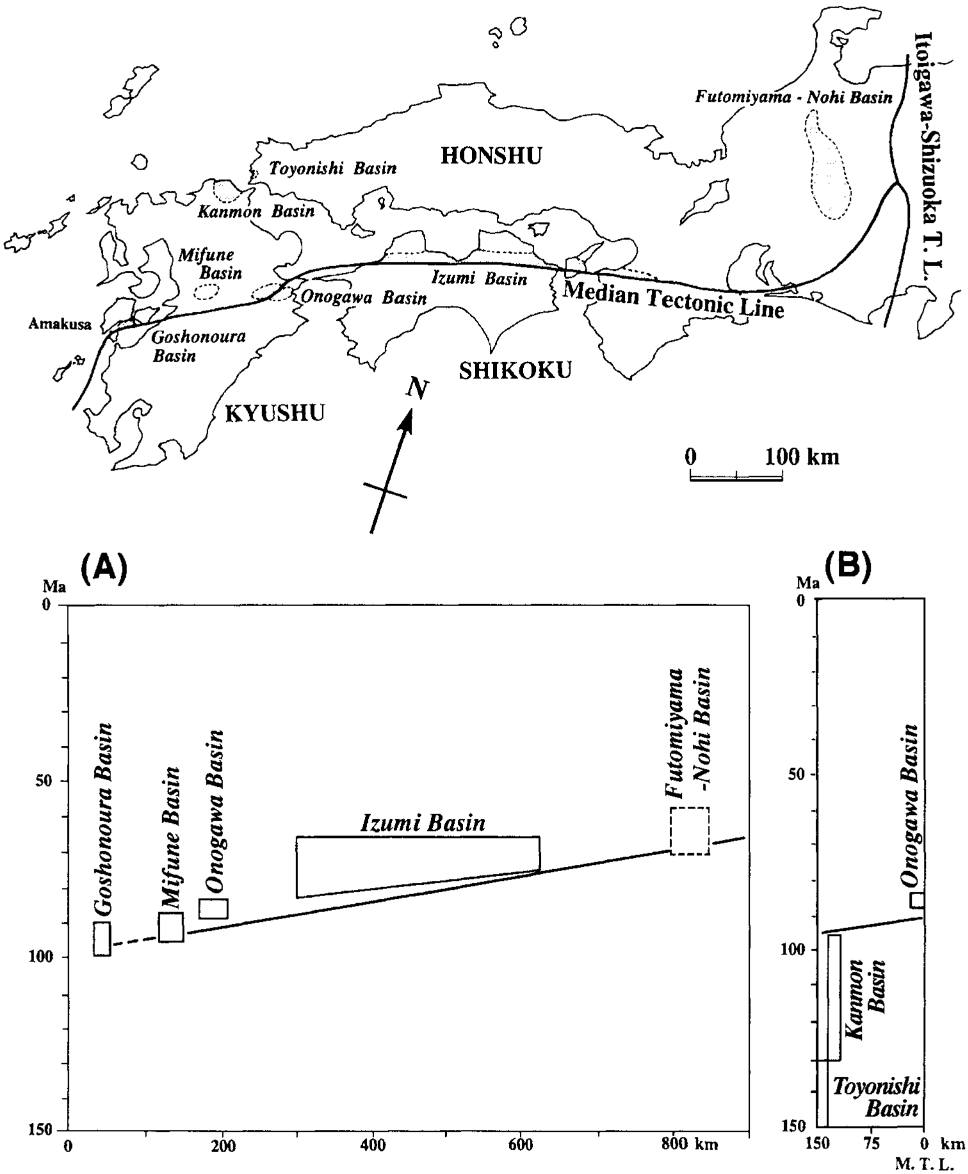

Figure 5. Distribution of the main Cretaceous basins in the Axial Zone of Southwest Japan (above) and the relation between magmatism and basin formation (below) along the Median Tectonic Line (A) and across the Line (B) (modified from Okada [90]). Data of magmatism are based on Kinoshita [86, 149].

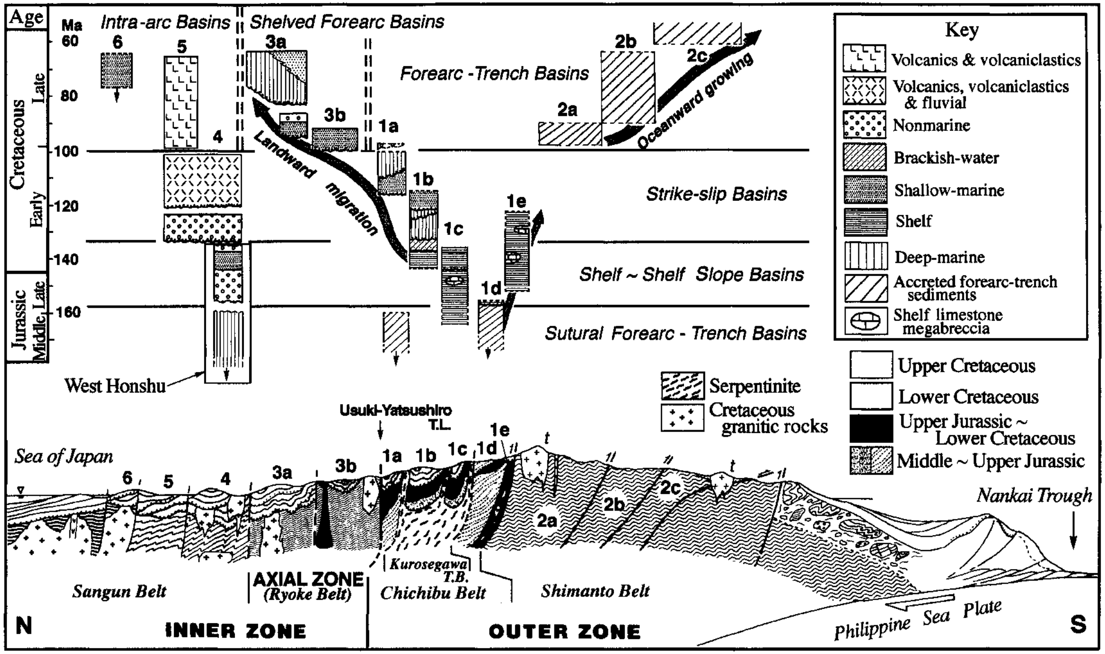

Figure 6. N-S profile of Kyushu Island showing the schematic structural features of Cretaceous sedimentary basins. Note a northward shift of sedimentary basins in the Kurosegawa Tectonic Belt and Axial Zone (modified from Sakai and Okada [73]). Not to scale.

1. Early Cretaceous basins in the Chichibu and Kurosegawa Tectonic Belts (1a: Yatsushiro and Tomochi Formations; 1b: Kawaguchi, Hachiryuzan and Hinagu Formations; 1c: Sakamoto and Uminoura Formations; 1d: Ebirase Formation; 1e: South Chichibu Belt); 2. Shimanto Belt (2a: Morotsuka Group; 2b: Makimine Group; 2c: Kitagawa Group), 3. Ryoke Belt (3a: Goshonoura and Himenoura Groups; 3b: Mifune Group); 4-6. Sangun Belt (4: Kanmon Group; 6: Late Cretaceous formations). t: Tertiary granitic rocks.

Albian Tomochi Formation (660 m) were deposited with northward shifting of their basins [73] (Figure 6). In the Kurosegawa Belt also, the start of sedimentation was retarded eastwards.

Post-Hauteriavian small and narrow Cretaceous basins in this belt with fairly thick clastic sediments show rapid subsidence, and rises of the basement repeated along strike-slip faulting such as the Kurosegawa Tectonic Line [73]. The basin formation shifted northwards [73]. These fault-bounded basins are analogous in origin to those in the California-type margin [73, 77].

2.4.2. Shimanto Belt

This is one of the most important tectonic units in the Japanese Islands (Figure 1). The Cretaceous sequence is retained in distribution to the northern subbelt, which extends over 1700 km from the northern Ryukyu Islands through South Kyushu, South Shikoku, the Kii Peninsula and the Akaishi Mountains, to the Kanto Mountains.

The strata in this subbelt are designated as the Lower Shimanto Supergroup, comprising flysch-type clastic sediments associated with small amounts of chert, limestone and pillowed basalt as accreted oceanic rocks [73]. They are lithologically subdivisible into two units: deep-sea fan turbidite facies and melange.

The strata are, in general, intensely folded and thrust, showing northward-dipping imbricate structures with ages younging southwards, and in part, weakly metamorphosed. These features are characteristic of typical trench accretionary prisms [73]. This lithologic feature is consistent throughout the Shimanto Belt. The total thickness of the Lower Shimanto Supergroup may reach more than 10000 m.

This accreted Cretaceous sequence in this belt in Kyushu is divided into three units, the Morotsuka (late Albian? or Cenomanian to Turonian), Makimine (Campanian to early Maastrichtian?) and Kitagawa Groups (late Maastrichtian? to middle middle Eocene?) and shows no accreted sediment in Early Cretaceous [73] (Figure 4). This means that the continental margin of Southwest Japan was under the transformed margin setting and no accretionary process took place during the Early Cretaceous (before the Cenomanian).

Megafossils are, in general, very rare [78], but radiolarian fossils are rather common. Thus, based on radiolarian biostratigraphy, structures and stratigraphy of the Shimanto Belt have been clarified in detail in Kyushu [73], Shikoku [79], Kii Peninsula [80], and the Kanto Mountains [81].

3. PHYSICAL ENVIRONMENT

The physical environment of the Japanese Cretaceous seems to have been controlled primarily by the change of plate movements. It was also for some time influenced by the plume magmatism. Such environmental features are summarized below mainly on the tectonic and sedimentologic basis.

3.1. Tectonic environment

3.1.1. Change of relative movements of oceanic plates

One of the important factors, which influenced the Cretaceous tectonic environment in the region between the Asian continent and the proto-Pacific realm, was changes in oceanic plate arrangements. It is generally believed that such changes in plate arrangements were envisaged by the Farallon Plate (-150 Ma), Izanagi Plate (150-85 Ma), Kula Plate (85-70 Ma), and Pacific Plate (70-0 Ma), among which the Izanagi and Pacific Plates subducted obliquely against the trench axis, whereas the Kula Plate nearly at right angles to the trench [82, 83]. Directions and movement rates of these plates were inferred by Seno and Maruyama [82, 83] (see Figure 4). The interactions of these plates including the kinematics of subduction of both oceanic plates and ridges with the Asian continent were constrained by the change of movement directions, as shown by the development of the Tan-Lu Fault system, variety of forearc basins, and igneous activity on the continental margin.

In this context, some workers insist that the subduction of volcanic ridges played an important role in generating temporal and spatial distribution of igneous rock types [84] and a large amount of igneous rocks [85, 86, 87]. Others consider that plume-related igneous activity, regardless of changes in plate subductions, controlled more importantly this geologic feature [88, 89, 90].

3.1.2. Tectonic framework of the pre-Cretaceous basement

The proto-Japan tectonic framework was established by the end of the Jurassic [91, 92, 93] by progressive accretionary processes of the Hida marginal and Sangun accretionary belts, the Permian accretionary belt, Triassic to Jurassic Tanba-Mino-Ashio accretionary belt, and Jurassic Chichibu accretionary belt around the Hida Belt as the protoconch of the Japanese Islands [92], among which the Chichibu Belt was incorporated by the collision of the Kurosegawa Tectonic Belt in the Jurassic time either as a micro-continent [93, 94] or as an island arc [95].

In the last decade, a strong similarity in radiolarian biochronology and sandstone petrography in addition to that in the lithofacies and geologic structures was revealed through extensive researches of these Jurassic accretionary belts. This means that both the belts, occupying different geotectonic positions in Southwest Japan from each other, were of the same Jurassic accretionary complex, and then they were juxtaposed as shown by the present duplicate distribution. Instead of a simple progressive accretion model, as mentioned above, modification of the tectonic framework has been examined in two different ways: the strike-slip model and the nappe model. Taira et al. [96] and Taira and Tashiro [97] showed a strike-slip model to explain a discordance in the distribution of the Late Mesozoic fauna and flora of the Japanese Islands and its adjacent regions by large-scale arc-parallel and left-lateral strike-slip faults. They insist that the intense shearing on the continental margin took place during the Early Cretaceous accompanied by the formation of local nappes and sedimentary basins.

It is very important to note that the Cretaceous sedimentary basins were developed on such a basement framework with many suture zones due to collisions and accretionary processes. These sutures are accompanied by serpentinite protrusions, which supplied plenty of chromian spinel grains to many Cretaceous basins [98, 99, 100] .

The pre-Cretaceous basement in Northeast Japan is divided into the Abukuma, South Kita-

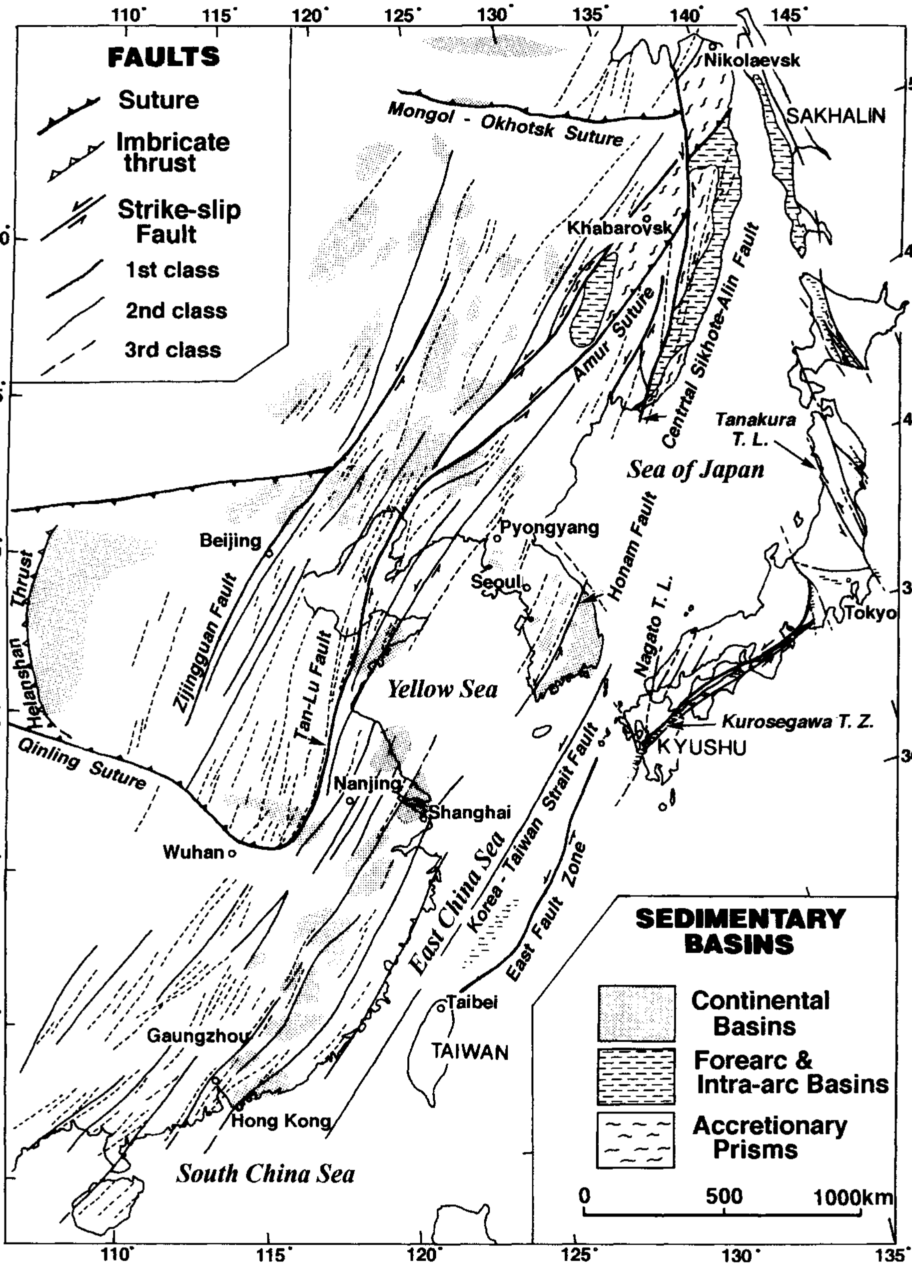

Figure 7. Map showing the distribution of Early Cretaceous sedimentary basins in East Asia and major strike-slip fault systems (modified from Sakai and Okada [73]).

kami and North Kitakami Belts in North Honshu and the Oshima Belt in Hokkaido. The basement rocks are composed basically of Permian to Jurassic accretionary complexes associated by Early Cretaceous granitic rocks [92] in a continental margin setting of the Eurasian Plate, on which small-scale epicontinental Late Cretaceous forearc basins were formed.

3.1.3. Tan-Lu Fault system and sinistral displacements

During the Early Cretaceous, the eastern marginal areas of the Asian continent were subject to strike-slip tectonism [73, 101] (Figure 7). This feature is manifest by a group of NEE-SWW faults with left lateral displacements over several hundreds kilometers [72, 73, 101]. This fault group is represented by the Tancheng-Lujian or Tan-Lu Fault system from Nanjing to Khabarovsk [73, 101, 102]. This fault system cuts older, E-W trending structures such as the Qinling and Mongol-Okhotsk Suture Zones, and was primarily active during the Early Cretaceous.

The NE- or ENE-trending fault zones in the East China Sea and Southwest Japan and the NW-trending fault zones in Northeast Honshu and Hokkaido represent the easternmost part of this fault system [73]. Some extensive strike-slip basins were formed along these faults. Thick-bedded nonmarine and marine sequences with fanglomerates or debris flow deposits started to accumulate at the base of the Cretaceous successions in both the Inner and Outer Zones of Southwest Japan. A marked change in sedimentation took place from eustasy-controlled shelf-alluvial conditions during the latest Jurassic to the earliest Cretaceous to the tectonics-controlled sequences in the late Early Cretaceous. This event, as indicated by an abrupt change of sedimentation with marked unconformity, is assigned to the Hauterivian. Such tectonic lines came into existence in the Early Cretaceous and controlled the pull-apart Late Cretaceous basins in the Axial Zone of Southwest Japan [103].

Many workers advocate that the Outer Zone of Southwest Japan was drifted from the initially southwestern geographical positions to the almost present position by the left-lateral fault activity represented by the Kurosegawa Tectonic Belt in the Early Cretaceous [104, 105]. Tashiro [104, 105] estimated the displacement of the Early Cretaceous basins in the Kurosegawa Tectonic Belt in Aptian and Albian as 500 to 1000 km based on the biogeographical analysis of molluscan fauna. On the basis of faunal and floral paleo-biogeography, Tazawa [48] also stressed the large scale left-lateral strike-slip motion along the Tanakura and Median Tectonic Lines in Early Cretaceous to Paleogene time, estimating the integrated displacement as large as 1500 km along those tectonic lines. Although main displacements of the Outer Zone ceased by the end of the Early Cretaceous, the left-lateral displacement of the Median Tectonic Line even in the Late Cretaceous time is estimated to have been at least 500 km [106].

Matsukawa et al. [52, 53] also reconstructed the paleogeography mainly based on nonmarine bivalve faunas in a different way, noting that the Chichibu Belt came into existence in late Early Cretaceous by migration from the south, and that the Sanchu Basin in the Kanto Mountains and the Katsuura-gawa Basin in Shikoku were formed in the forearc and the Kanmon and Upper Tetori Basins in the intra-arc regions, respectively.

In Northeast Japan also, sinistral strike-slip fault movement was active during the Cretaceous, developing the Tanakura Tectonic Line in the Abukuma Belt with displacement of more than

240 km, the Hizume-Kesennuma Fault (before 110-120 Ma) with a displacement of 40 km in the South Kitakami Belt, and the Taro Fault (121-112 Ma) in the North Kitakami Belt [107].

3.1.4. Tectonic change of the continental margin

Cretaceous sedimentary basins in and around of the Japanese Islands developed in responce to the tectonic changes along the continental margin of East Asia. Most of them, however, represent a hybrid nature reflecting a wide variation of basin types. In Southwest Japan, the evolution of the Cretaceous continental margin defines three types and stages; the passive margin (latest Jurassic to earliest Cretaceous), transformed margin (late Early Cretaceous), and oblique-slip convergent margin (Late Cretaceous). These three stages are delineated by two marked tectonic events, which took place at the late Early Cretaceous and the earliest Late Cretaceous time.

3.1.4.1. Early Cretaceous event

There is no evidence to indicate the accretionary-subduction processes during the latest Jurassic to the earliest Cretaceous. The sedimentary basins exposed separately in the Outer and Inner Zones of Southwest Japan are composed of clastic sequences, in which orthoquartzite clasts and quartz-arenaceous sediments accumulated during the lowstand stages. The sequence evolution of these basins was primarily controlled by eustatic sea-level change. These basins developed on more stable continental shelf with older continental basements. The Lower Cretaceous chaotic deposits exposed to the south of the Jurassic accretionary complex correspond to a shelf-slope component in this period. The nature of these sedimentary basins from the shelf to slope regions suggests that the tectonic setting at the eastern margin of the East Asia continent was a passive margin.

After the ephemeral development of the eustasy-controlled shelf basin during the earliest Cretaceous, the eastern part of the East Asian continent turned into a transformed margin at the late Early Cretaceous. This event was preceded by an oblique-slip convergent margin setting until the early Late Jurassic [81, 108]. The tectonic inversion in this period is documented well in abrupt facies changes with a marked unconformable contact between the lowest and Lower Cretaceous formations in both the Inner and Outer Zones of Southwest Japan.

The late Early Cretaceous also lacked subduction-related components. Many pull-apart basins accompanied closely by the arc-parallel shearing became to appear at the beginning of the late Early Cretaceous. In Southwest Japan, these pull-apart basins occur along some suture zones, such as the Hida Marginal, Nagato and Kurosegawa Tectonic Belts. An unconformity with thick debris-flow deposits and abrupt changes in depositional system between the earliest Cretaceous and the late Early Cretaceous sequences are also indicative of tectonic events [73]. The development of strike-slip basin system in this period coincided with the main phase of shearing in the eastern part of the Asian continent [101]. In the synchronism with and similarity to the intra-continental basins associated with the Tan-Lu Fault system [73], the late Early Cretaceous basins in Southwest Japan are grouped into a component of the eastern part of the Tan-Lu Fault system.

The Early Cretaceous shallow-marine and deep-marine basins in the Kurosegawa Tectonic Belt are considered to have been formed as pull-apart (or a sutural forearc) basins on the semi-continental blocks (the Kurosegawa Massif). The sequential evolution of these basins records the suturing and tectonic amalgamation between the proto-Southwest Japan Arc and the suspect massif, which had been accreted in a certain time of the Jurassic. According to Seno and Maruyama [82], the Izanagi Plate had converged diagonally against the East Asian continent. This oblique subduction should have caused substantial left-lateral slip within the crushed Kurosegawa Massif.

It is still poorly understood why the Tan-Lu Fault system could have developed in such a wide area of the East Asian continent: over 4000 km long and 1000 km wide. Intra-continent and continental margin type magmatism, which took place on more inner side of the Asian continent, seems to be important. Takahashi [84] proposed a ridge subduction model to explain such magmatism. Tectonic situation of the East Asian continental margin in this period was similar to that of the California-type margin, where the shallow subduction of oceanic plates with active ridge had continued since the Oligocene [109, 110]. A similar transformed margin can be expected for the tectonic setting of the East Asian continental margin during the late Late Cretaceous [73].

The Tan-Lu tectonic event is supported by many lines of paleo-biogeographical evidence indicating a large-scale lateral displacement of continental margin [97, 104], formation of regional nappes [111, 112] and flower structures [73], and serpentinite protrusion and related sediments [113, 114]. The fundamental tectonic framework characterized by the duplication of the Jurassic accretionary prism in Southwest Japan is, therefore, assumed to have been established by the end of the Early Cretaceous.

3.1.4.2. Mid-Cretaceous event

Coexistence of the felsic-igneous activity and forearc-trench basins characterizes the entire East Asian continental margin during the Late Cretaceous. This implies that the Late Cretaceous continental margin of East Asia was placed in a typical Cordilleran-type margin setting. The timing of this tectonic change from the transformed to convergent margins occurred at the mid-Cretaceous.

Early Late Cretaceous shallow-marine basins, such as the Mifune and Goshonoura Basins in West Kyushu and the East China Sea Shelf Basin, were formed on the upper shelf, and show a fault-controlled half-graben profile. They correspond in age to the deep-marine forearc-trench basin in the lower part of the Lower Shimanto Supergroup. The Campanian to Maastrichtian Onogawa and Izumi Basins, on the contrary, are typical pull-apart basins resulted from the left-lateral movement of the Median Tectonic Line, which is related to an oblique-slip subduction. Spatial and temporal distribution of subduction-related sedimentary basins of various forms are characteristic of the eastern margin of the Asian continent. This variation and discontinuity may be related to differences in the direction of oceanic plates and ridges consumed as well as in the structural framework of arc basements.

The mid-Cretaceous tectonic inversion is reflected in basin shifts, marked lithologic change with unconformities, and abrupt basin expansion. Among them, basin expansion with tilting was revealed in many basins in the Tan-Lu Fault system; for example the Songliao [115], Subei-South Yellow Sea Basins [116], and the East China Sea Shelf Basin [117]. These basins are referred to as rift or rift-related basins [117, 118, 119]. Basin expansion in this region is related closely with the change of paleostress field along the Tan-Lu Fault system. Transtension followed by transpression may have been especially significant in the development of rift basins. The tectonic inversion due to the change of paleostress field at this time was recognized in some regions of the Tan-Lu Fault system [118, 120, 121, 122], and in the regions within the Qinling Suture and its extension (the Baiwan Basin and the northern Jiangsu Basin [123]).

As typically represented by the Shimonoseki Subgroup in Southwest Japan and the Hayang Group in South Korea, intense igneous activity within the sedimentary basin was a remarkable feature of the mid-Cretaceous event. Although igneous activities are common in the western basins of China [124], it is debatable that the initiation of oceanic plate subduction resulted in the formation of the widespread extensional basin and igneous activity so far from the trench. As pointed by Yano and Wu [88, 89] and Okada [90], the mantle-related rifting seems to be an alternative possible mechanism to account for the genesis of widespread extensional basins in the East Asian continent. In this respect, Okada [90] proposed plume-related basin formation.

3.1.5. Tectonic setting of sedimentary basins and their origin

Most of the Late Mesozoic basins in East Asia are, in general, polygenetic in origin, as pointed out by Li [117], Hsu [119], and Liu [121]. Due to the tectonic inversion, their present structural profiles do not always show an initial structural feature of the continental margin.

According to tectonic types, sedimentary basins in East Asia are divided into subduction-related basin, pull-apart basin, and rift basins [90], corresponding roughly to trench-forearc basins, intra-arc basins and continental basins, respectively.

3.1.5.1. Subduction-related basins vs. trench-forearc basins

Subduction-related basins are subdivided into the trench basin, trench-slope basin and fore-arc basin, all of which are formed in close relation to subduction of oceanic plates. Because the facies from such basins are difficult to distinguish from each other, they are grouped as trench-forearc basins for convenience. They are represented by the late Tithonian to Albian Sorachi Group and the Aptian to Maastrichtian Yezo Supergroup of the Yezo Arc-Trench System in Hokkaido, and the Albian to Eocene Lower Shimanto Supergroup in Southwest Japan.

In Northeast Japan, Hokkaido Island was separated into two tectonic masses: the Yezo Arc-Trench System to the west and the Okhotsk Paleoland to the east [10]. Kimura [32] regarded this Okhotsk Paleoland as a proto-Kurile Arc, because no continental crust is encountered beneath the paleoland. These two masses were separated by the Sorachi Ocean, a part of the Izanagi-Kula Plate, which subducted beneath the two masses. Thus, trench-forearc basins were developed in the Yezo Arc-Trench System, which were deposited by the Yezo Supergroup [10]. The Okhotsk Paleoland was also surrounded by trench-forearc basin deposits such as the Campanian to Eocene Nemuro Group on the south side [103] and the Tithonian to Albian Nikoro Group, Campanian to Paleocene Saroma and Yubetsu Groups in the Tokoro Belt on the west side [103].

In the Outer Zone of Southwest Japan, subduction complex is represented by the Lower Shimanto Supergroup, in which the Morotsuka (Albian? to Turonian), Makimine (Cenomanian to early Maastrichtian) and Kitagawa Group (late Maastrichtian? to middle middle Eocene?) are successively discriminated in South Kyushu. They show a large-scale imbricate stacking of strata with their ages younging oceanwards. Thus, from the arc side to the trench side, early Late Cretaceous, Late Cretaceous and Latest Cretaceous to Early Paleogene subduction complexes were added to the accretionary prism. This means that, during the Early Cretaceous, no significant accretion took place along the oceanic plate margin of Southwest Japan (Figure 4). Instead, there must have been a transform or oblique slip contact between them.

A similar phenomenon was observed by Takahashi and Ishii [81, 108] in the Early Cretaceous in the Chichibu Belt of the Kanto Mountains, where chaotic deposits indicate a tectonic change from convergent plate margin to oblique-slip nature in the earliest Cretaceous time. This situation seems to have been different in the regions north of the Tanakura Tectonic Line, where both Early Cretaceous accretion and volcanism proceeded together (Figure 4), though more detailed chronological data are needed for accreted rocks and magmatism.

3.1.5.2. Pull-apart basins

In Southwest Japan, particularly in Kyushu and Shikoku Islands, pull-apart basins are common in forearc to intra-arc regions [73]. They are generally fault-bounded trough-like depressions arranged in parallel to tectonic lines, and are separately scattered, small, elongated, and filled with clastic sediments of nonmarine, paralic to shallow-sea and deep-sea facies [72, 73, 103] of substantial thicknesses.

Pull-apart basins are characteristic of forearc and intra-arc regions in the Sanchu and Choshi Basins in central Honshu [52, 53, 125], and the Kurosegawa and Axial Belts in Kyushu [72, 73, 103], as well as of continental regions including the Yongdong Basin and related basins in South Korea [126, 127], the Yilan-Yitong Fault-related basins east of the Songliao Basin, and elongated basins along the Tan-Lu Fault east of the Hefei Basin in North China. In general, they show half-graben profiles. Miyata [128] demonstrated experimentally that the scale of graben structures is controlled by the degree of slope angles of main strike-slip faults.

In the Kurosegawa Belt, basins small in size, ranging from 10 to 50 km and short in duration about 5 to 15 m.y., show a northward migration [73] (Figure 6). A mechanism of this basin migration is explained in a way that the strength of coupling of different terranes beneath arc margins increased progressively arcwards, which formed depressions along the coupling sutures.

Relations between the development of pull-apart basins and magmatism is clear in intra-arc basins along the Median Tectonic Line in Kyushu and Shikoku Islands. In these areas, small Cretaceous basins with thick clastic sequences are exposed (Figure 5). The westernmost is the latest Albian to earliest Turonian Goshonoura Basin, the western central Mifune Basin from the Cenomanian to Turonian, the eastern central Onogawa Basin from the Turonian to Santonian, and further in the east the Campanian to Maastrichtian Izumi Basin in Shikoku Island. The Izumi Basin itself extends over 500 km, started its sedimentation in the Campanian first in the west and then in the Maastrichtian in the east. The Onogawa and Izumi Basins are characterized by extraordinarily thick piles of turbidites [103].

It is worthwhile to point out that the depocenters in these basins shifted gradually eastwards and got younger in the same direction along the Median Tectonic Line [3, 103] (Figure 5). This diachronism in basin formation is very similar to the age trend in magmatism in Southwest Japan (Figure 5) with a rate of about 1/30 m.y./km [86]. This rate is close to that of movements of the Kula and Pacific Plates [82, 83].

Therefore, it is clear that formation of pull-apart basins and subduction-related magmatisms along the Median Tectonic Line are closely related to each other. At the same time, depocenters of the basins in the western Axial Zone migrated northwards, as in the Kurosegawa Belt.

3.1.5.3. Back arc and continental basins vs. rift basins

Back arc and continental basins are characterized mainly by rifting nature, which is typical, in particular, of the continental region from the Korean Peninsula to China. They are, thus, mostly defined tectonically as rift basins.

The continental nature of the Japanese Islands was well developed during the latest Late Jurassic to Early Cretaceous time, when the tranform continental margin appeared. The Tetori and Kanmon Basins represent such a tectonic condition, both of which are bounded by the Tan-Lu Fault system.

Rift basins in East Asia are generally characterized by their large scale and fault-bounded structures. A typical example is the Valanginian to Cenomanian Kyongsang Basin in the Korean Peninsula, measuring about 250 km long and 125 km wide. It shows a half-graben structure tilted eastwards [72].

The East China Sea Shelf Basin is also characterized by composites of fault-controlled half-grabens formed on the Proterozoic basement. They are narrowly elongated en echelon with NE to NEE trend, extending more than 1000 km. Most of the basins are filled with Upper Cretaceous, Paleocene and Eocene sediments nearly 4000 m thick [122]. All these half-grabens initiated in the Late Jurassic, but major subsidence took place in the Late Cretaceous.

3.2. Sedimentary environment

3.2.1. Basin characters and environments

3.2.1.1. Forearc basins

Trench-forearc basins on the Japanese Islands are generally characterized by a subduction complex composed of chert-greenstone and deep-sea fan clastic turbidites, as are well recorded in the Outer Zone of Southwest Japan and in the Kamuikotan, Hidaka and Tokoro Belts in Hokkaido.

Trench deposits are also found typically in the Shimanto Belt in the Outer Zone of Southwest Japan. They are characterized by highly deformed strata originally deposited in trench to forearc basins typical of trench subduction areas [129, 130]. The Cretaceous sequence of the Shimanto Supergroup is composed mainly of alternations of sandstone turbidites and shale, and subordinately of chert, red shale, micritic limestone and pillow basalt, attaining the whole thickness of more than 10000 m [103].

In Hokkaido, deep-sea trench environments are indicated by the late Tithonian to Albian Sorachi Group of the Kamuikotan Belt, Early Cretaceous to Eocene Hidaka Supergroup in the Hidaka Belt for the Yezo Arc-Trench System and the Late Jurassic to Late Cretaceous Nikoro

Group in the Tokoro Belt and the Campanian to Eocene Nemuro Group in the Nemuro Belt for the Okhotsk Paleoland. Among them, the Nikoro Group consists of sea-mount rocks and surrounding deposits initially formed on the Izanagi-Kula Plate at the Equator region [13].

Forearc environments are shown by the Yezo Supergroup in the Ishikari and Kamuikotan Belts for the Yezo Arc-Trench System and the Saroma and Yubetsu Groups in the Tokoro Belt. The former shows a spectrum from littoral, shallow-marine and offshore, to trench-slope environments [10, 131, 132].

Lower Cretaceous deposits of the Choshi and Sanchu sequences in the Kanto area and those in the Katsuuragawa area in East Shikoku are of the forearc [52]. Many small patches of the Upper Cretaceous in the Abukuma and Kitakami regions are also forearc deposits of alluvial-fluvial and shallow marine facies [132, 133].

The forearc basins in Southwest Japan are developed in the Chichibu Belt. Particularly in Kyushu, they are found in the Axial Belt, the southern part of the Ryoke Belt, and the Kurosegawa Belt. These basins consist of nonmarine to shallow-marine and deep-marine sediments. The former is represented by the Kawaguchi, Kesado and Yatsushiro Formations in the Kurosegawa Belt and the Goshonoura, Mifune, and the lower sequences of the Himenoura and Onogawa Groups, which are characterized by river-mouth bar and lagoonal facies with abundant molluscan and floral fossils. The latter, deep-marine basins are represented by the Uminoura, Hachiryuzan, Hinagu, Tomochi Formations, and the upper sequences of the Himenoura and Onogawa Groups with rhythmically interbedded turbidite facies suggesting deep-sea environments.

3.2.1.2. Intra-arc and back-arc basins

The northern subzone of the Inner Zone of Southwest Japan is characterized by relatively large sedimentary basins of an intra-arc type. They include the Lower Cretaceous Kanmon Group, and the Upper Cretaceous Shunan, Hikami and Abu Groups. The latter three are the basin-fills of cauldrons in the magmatic intra-arc region [134, 135, 136]. The Kanmon Group consists of the fluvio-lacustrine Wakino Subgroup in the lower and the alluvial and volcanogenic Shimonoseki Subgroup in the upper [1, 137]. The Shunan, Hikimi and Abu Groups are characterized by felsic volcano-plutonic rocks [134, 135].

No clear back-arc basin is identified in the Japanese Islands during the Cretaceous.

3.2.1.3. Continental basins

A typical example of continental basins juxtaposed to the Japanese Islands is the Lower Cretaceous Kyongsang Basin in South Korea. It is filled with sediments of fluvio-lacustrine environments, as shown by frequent occurrence of channel beds and argillaceous lacustrine limestones [71]. The occurrence of caliche, calcareous nodules and nodular marl indicates seasonal aridity on the floodplain environment [70].

3.2.2. Red-bed problem

Cretaceous red beds are widely distributed in East Asia. They are divided into two groups: with or without evaporites [138, 139]. The evaporite-bearing red beds seem to occupy the basins developed from Jurassic salt-lakes [139]. The non-evaporitic red beds are distributed in the Kyongsang Supergroup in South Korea and Cretaceous to Paleogene strata in Southwest Japan.

The occurrence of red beds implies in general alternating hot/dry and/or hot/humid climate [139]. Such a climate not only causes deep weathering of source rocks rich in mafic silicates, but also the oxidation of Fe-bearing minerals to form hematite pigmentation on detrital grains, giving sediments red coloration. Miki [138], comparing chemical features of red beds between the Hong Kong Cretaceous and the Korea and Kyushu Cretaceous, showed that the Hong Kong red beds with extremely high values of Fe^2O^3/FeO were deposited under drier conditions than the Korean and Kyushu red beds, which are similar to each other in chemistry. In this respect, the distribution of red beds in the Japanese Islands is worthwhile to be specially mentioned.

Red beds in the Japanese Islands are mainly found in the Inner Zone of Southwest Japan, which are grouped into the red-bed petroprovince by Okada [74], although some are found sporadically in the Lower Cretaceous shallow-marine to littoral sediments in the Outer Zone of Kyushu and Shikoku Islands as well as in the Lower Cretaceous shallow-marine to littoral deposits at Miyako in Northeast Japan and in the Upper Cretaceous littoral to shallow-marine sediments of the Mikasa Formation of the Middle Yezo Group and the Hakobuchi Group in the Ishikari Belt of Hokkaido.

Tashiro [105] advocated that the displacement in distribution of red beds between the Inner Zone and the Outer Zone of Southwest Japan indicates a tectonic movement over the distance of more than 300 km due to the left-lateral displacement of the Median Tectonic Line during the Late Cretaceous. He further pointed out that such remarkable dislocations of the Outer Zone are reinforced by faunal and floral evidence [105]. On the contrary, Matsukawa et al. [52, 53] explained in a different way that the Chichibu Belt took its present position before the Early Cretaceous, where warm and cold currents controlled faunal distributions recognized by Tashiro [105].

The main reason why Cretceous red beds in the Japanese Islands are restricted in distribution to the Inner Zone of Southwest Japan is that this region was either on land or under shallow-sea environments during dry/humid climate under high temperatures during the Cretaceous. Even in other areas under similar environments, red beds should also have been developed.

3.2.3. Influence of a bolide impact

A global effect of a large meteorite impact at the interval of the K/T boundary is also observed in the Nemuro Belt of Hokkaido [140]. A 6-10 cm thick layer of pyrite-rich, fossil-poor black claystone was detected by Saito et al. [140], though no iridium enrichment was observed. They also found sudden changes in terrestrial palynomorphs in this interval, and interpreted that a devastation of the land flora was caused by a catastrophic event such as a meteoritic impact. Further regional influence of this bolide impact in other areas in the Japanese Islands has not been made clear.

3.3. Paleogeography and topography

There is no doubt that the Inner Zone of Southwest Japan was connected with the Asian continent somewhere to the east of the Korean Peninsula during the Cretaceous [141, 142]. This situation persisted until the opening of the Sea of Japan at 28 Ma [143]. Otofuji and Matsuda [143] showed an original position of Southwest Japan before the rifting event of the Sea of Japan. On this line of evidence, some attempts to reconstruct the paleogeography during the Cretaceous have been made, for example, by Matsukawa and Tsuneoka [144] for the Berriasian, by Tashiro [105] for the Hauterivian-Barremian and for the Albian-Cenomanian, based on the analysis of molluscan faunas, by Kimura [141] for the Late Cretaceous, and by Maruyama et al. [142] in a series of paleogeographic maps from 750 Ma to the present.

The topography of the Japanese Island arc during the Cretaceous must have been quite high and under intense erosion. As reviewed by Okada [145], the Cretaceous sequence in the Japanese Islands shows anomalously high sedimentation rates of clastics even during the local transgression epoch called the Urakawa Transgression in Cenomanian to Campanian, which corresponds to the global high sea-level [146]. The sedimentation rates of Cretaceous units in the forearc region, for example, are 153 m/m.y. for the Goshonoura Group, 167 m/m.y. for the Mifune Group, 2857 m/m.y. for the Onogawa Group, and as high as 4000 m/m.y. for the Izumi Group.

These high sedimentation rates correspond as a modern analogy to those in the Himalayas, Taiwan and the Akaishi Mountains in central Honshu, where sedimentation rates of 30000 m/m.y. for the Middle Siwalik sequence of the Himalaya Foreland, about 1000 m/m..y. for the Pliocene of Taiwan, and about 1700 m/m.y. of the Pleistocene in the Akaishi Range area [145]. Such high sedimentation rates reflect very rapid upheaval as much as 3-4 mm/y for the Himalayas, 2-5 mm/y for the northern coastal range of Taiwan, and 4 mm/y for the Akaishi Range are obtained [145]. Sugai and Ohmori [147] have given the denudation rate of 1.07 and 1.02 mm/y for the present Akaishi and Kiso Ranges, respectively.

This suggests a similar tectonic framework during the Late Cretaceous in the Japanese Islands to that during the Quaternary in Japan and Taiwan. This further means that tectonically active uplifted areas were subject to intense weathering and erosion due to high precipitation accompanied by active transportation.

3.4. Episodic supply of clastics

As mentioned above, a huge supply of clastics was encountered both in Southwest Japan and in Northeast Japan, which resulted in episodic growth of accretionary complexes. In Southwest Japan, two major events of clastic supply are recognized in early Early Cretaceous and in Late Cretaceous times (Figure 4), which roughly correspond to the northwestward to westward motion periods of the proto-Pacific plates. Late Cretaceous supplies were enormous both in the intra-arc and forearc regions [145], as mentioned above.

In Northeast Japan, clastic supply was rather continuous from Early Cretaceous to earliest Paleogene, as is well documented in Hokkaido (Figure 4). In particular, this trend was accentuated by the deposition of enormous amounts of clastics of the Late Cretaceous to Paleocene Nakanogawa Group, which corresponds to the westward to northwestward period of motion of the Kula Plate.

Kimura [32] related the rapid supply of clastic sediments into the trench in Northeast Japan to the mountain building along the Mongolia-Okhotsk Suture Zone. A similar situation with the rapid supply of huge amounts of clastics to the Chichibu and Shimanto trenches has been assumed by Sakai [148] for the Qinlin Suture Zone in Central China. Kimura [141] also discussed the epsodic and rapid growth of the Japanese Islands during the Late Cretaceous, suggesting very young oceanic crust and/or ridge subduction.

4. CONCLUDING REMARKS

Knowledge of the Cretaceous geology of the Japanese Islands has been enhanced by abundant new data gathered in the last decade. This has enabled the recognition of (1) the arc-parallel strike-slip nature of many tectonic components; (2) a non-accreted span at a certain segment along the continent-ocean boundary in the Early Cretaceous time; and (3) an extraordinarily high rate of clastic sedimentation during the most of the Cretaceous Period. It is suggested that the second condition may have generated the plume-related igneous activity at the East Asian continent. The third condition reveals the significance of the uplifted belts of the Qinling Suture and the Mogolia-Okhotsk Suture as the major suppliers of a huge amount of clastic materials. These factors seem to have rendered great deal of influence on both physical and biological environments during the Cretaceous.

Further study is needed for the delineation of paleogeography of the Japanese Cretaceous at the stage level.

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Changes in Cretaceous ammonoid diversity and marine environments of the Japanese Islands

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This study surveys 902 species of Cretaceous ammonoids in Japan, and identifies species diversity changes during 29 Japanese Cretaceous substages. The species diversity fluctuates between the peak of 92 species in upper Albian and the lowest of less than 40 species in lower and middle Albian, upper Cenomanian, lower Coniacian and upper Campanian onward. More than half of these periods of the lowest diversity except that in the lowest and uppermost Cretaceous, correlate with oceanic anoxic events/subevents, indicating that species diversity of the Japanese Cretaceous ammonoids was primarily controlled by such oceanic anoxic events/subevents.

1. INTRODUCTION

As described by Okada and Sakai [1], the Cretaceous marine sediments are widely distributed in the Japanese Islands, and yield many ammonoids, which are critical for international correlation[2]. The frequent occurrence of inoceramid bivalves is also important for correlation within Japan and neighboring parts of the Far East. The chronostratigraphy is based on

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planktic and benthic foraminifers, radiolarians, nannofossils, paleomagnetism and radiometric dating in addition to ammonoid and inoceramid data [3, 4, 5], which gives greater bio stratigraphic resolution to the Japanese ammonoid zonation. The mid-Cretaceous oceanic anoxic event (OAE 2) at the boundary between the Cenomanian and the Turonian Stages (OAE 2) is consistent with this biostratigraphy [6, 7].

Use of high resolution global events like OAE [8] is helpful in tying the Japanese sequence, and although OAE 2 has been studied in detail, the details of OAE 1 remain imprecise [9].

The oceanic anoxic event in the Coniacian and the Santonian Stages (OAE 3) has not yet been clarified. Causal mechanisms for the occurrence of anoxic to dysoxic waters have been proposed, but as yet without consensus [8].

We have surveyed all Cretaceous ammonoids from Japan, and have analysed diversity changes of ammonoids at the species level. Ammonoids are known to have been influenced by the environmental changes, for which the term bioseismometer was introduced by House [10]. We would be able to expect to clarify the environmental changes of Japan and the Northwestern Pacific as well through the study of ammonoids. Thereupon, we describe the species diversity changes of ammonoids and discuss the relation mainly with oceanic anoxic events.

2. AMMONOIDS AS BIOSEISMOMETERS

House [10] showed biodiversity changes in ammonoid families at a scale of 2 m.y. from the Devonian to the end of the Cretaceous. In his analysis of Cretaceous ammonoid families, the high biodiversity peaks occur in the Barremian, Albian and Cenomanian, and the minima occur in the Berriasian and Aptian. Origination of new families ceased in the Santonian. Biodiversity changes of ammonoid families correlate significantly with the sea-level changes throughout ammonoid history.

Wiedmann [11] reported generic diversity changes at a scale of stage, and noted that diversity increased at the time of transgression and decreased at the time of regression, based on the knowledge of the sea-level changes at the time. Kennedy [12] overprinted the generic diversity changes on the figure of the area of epicontinental sea, and showed visually the significant correlation between them. The recent work of House [10] took 2 m.y. as the time unit and sea-level change data from Haq et al.[13], and statistically compared them with the result that the correlation between ammonoid diversity and sea-level changes is significant.

Sepkoski and Raup [14] studied 3500 protozoa, invertebrates and vertebrate in the 82 stages younger than the Paleozoic Leonardian. Their analysis for the twelve stages of the Cretaceous shows the largest extinction rate in the Cenomanian and the Maastrichtian. Sepkoski [15] shows biodiversity changes at the generic level, with high extinction rates in the Aptian and the Cenomanian. Sepkoski and Raup [13, 16] reported that the extinction rate increased progressively from the Aptian through the Albian to the Cenomanian, and that in the Campanian and Maastrichtian very high extinction rates occurred. In contrast, the net family loss reported by House [10] occurred in the lower Aptian, lower Turonian and upper Campanian. Comparisons between their results are complicated by differing metrics, but the trends seem to be similar except the uppermost Cretaceous. Both results suggest two comparatively larger scale of extinctions in Aptian and upper Cenomanian.

3. SPECIES DIVERSITY CHANGES IN THE JAPANESE CRETACEOUS AMMONOIDS

We have surveyed records of all Cretaceous ammonoid species (N=902) from Japan in order to analyze the species diversity changes of the Japanese Cretaceous ammonoids. In Japan 29 substages are used for the Cretaceous [4, 5] in addition to the European standard stages and substages. For instance the lower substage of Cenomanian is divided into two domestic substages, K4a1 and K4a2.

The analysis of the species diversity change for 902 Japanese Cretaceous ammonoids shows high diversity in lower Barremian, upper Aptian, upper Albian, middle and upper

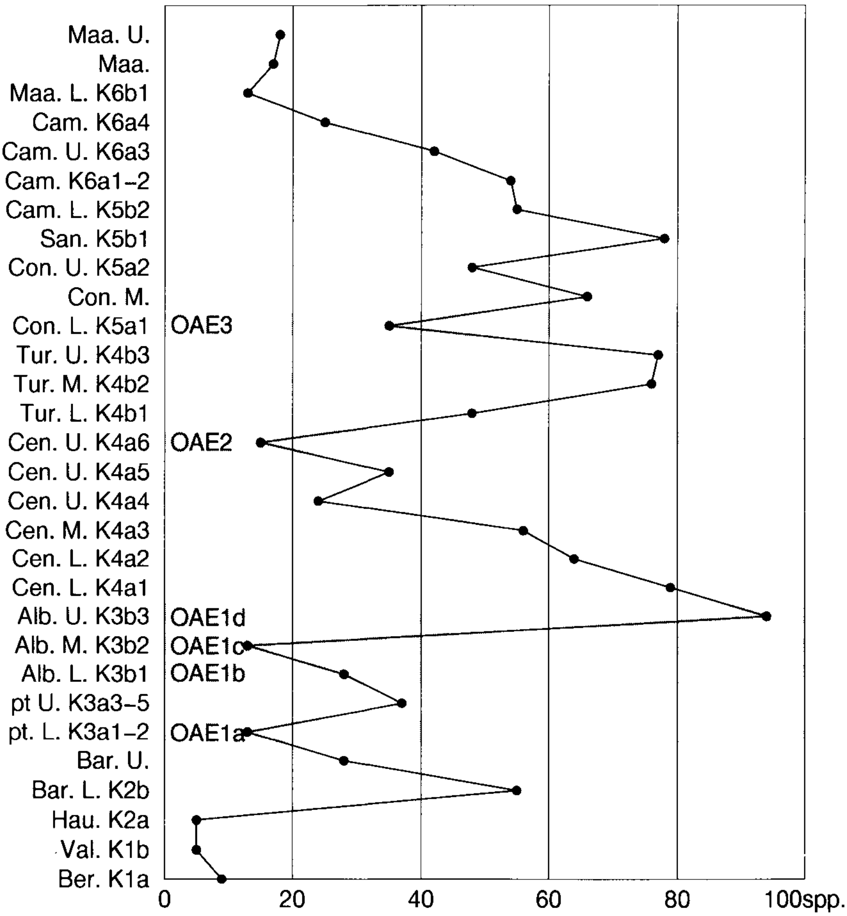

Figure 1. Species-diversity changes in Cretaceous ammonoids of Japan. The horizontal line shows the number of ammonoid species. K1a to K6b are letter nominations of the Japanese Cretaceous substages (For more details, see the charts of Toshimitsu et al. [4, 5]).

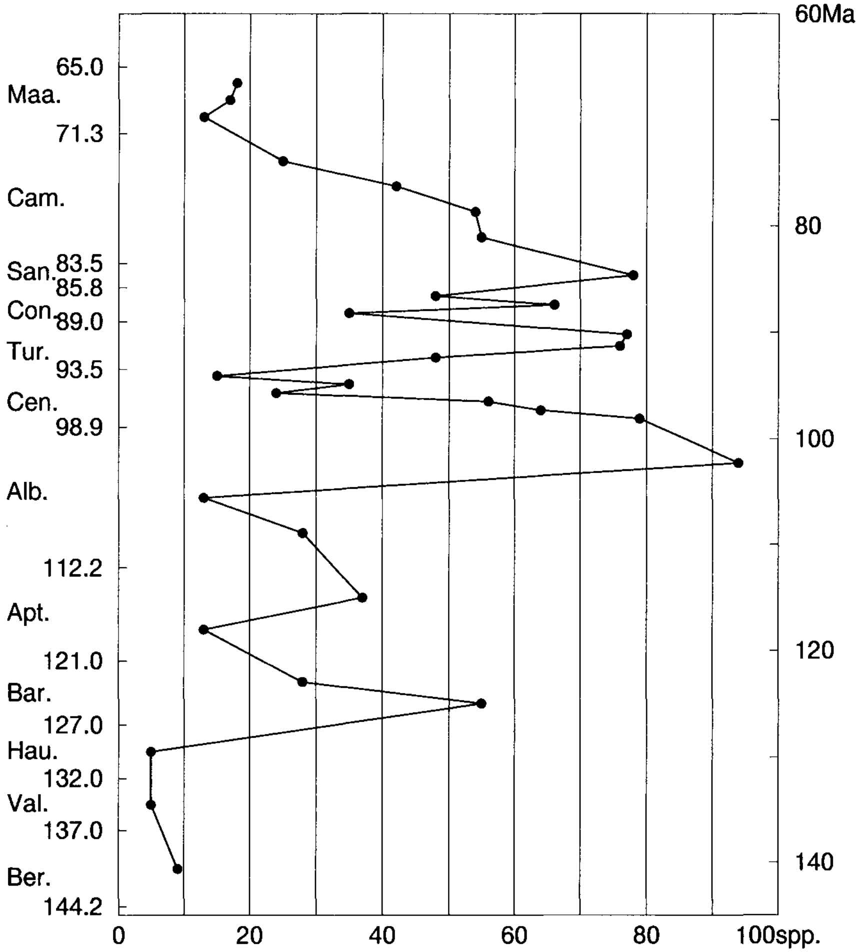

Figure 2. Species-diversity changes in Cretaceous ammonoids of Japan on a radiometric scale.

Turonian, middle Coniacian and Santonian stages with the highest diversity in upper Albian Substage (Figure 1). It is natural that there are minima of diversity between some peaks as a pattern. The minima occur in lower Aptian, middle Albian, and upper Cenomanian Substage. The analysis also shows that the diversity increased comparatively rapidly and decreased slowly (Figure 2). In the Upper Cretaceous, diversity decreased after the peak in the San tonian.

House's [10] minima occur in the Berriasian, upper Aptian and upper Albian, and the progressive decrease is in Cenomanian onward. In the species diversity analysis, a minimum is also detected in Berriasian, but there are differences on the scale of substage for Aptian and

Albian. A minimum is detected in Coniacian in the analysis of the species diversity changes, but not in the family diversity changes.

The net family loss of the family diversity changes shows larger absolute values in Lower Aptian, Lower Turonian and Upper Campanian. In the turnover rate of species diversity changes, more numerous minima are detected, which do not well match the results of the net family loss.

4. RELATION OF SPECIES DIVERSITY CHANGES WITH MARINE ENVIRONMENTAL CHANGES

Changes in species diversity during the Cretaceous have been attributed to the evolution of the predators such as teleostean fish and crustaceans [17], and, alternatively, the selection pressure in relation with the sea-level changes [18]. Stanley [19] identified temperature, salinity, space, dissolved oxygen, food resources, and struggle in addition to predation as limiting factors for marine animals. House [10] noted the significant correlation between the diversity changes of ammonoid families and the sea-level changes. Here, we correlate ammonoid species diversity changes with oceanic anoxic events. Three oceanic anoxic events during Cretaceous are recognized as OAE 1 to 3 [20]. The Lower Cretaceous OAE 1 is divided into four subevents, OAE 1a to 1d [21]. Although OAE 2 is recognized as global, OAE 1 and 3 have not been shown yet to be as widespread. Hirano and Fukuju [22] and Takahashi et al. [23] distinguish four subevents of OAE 1 in the sections of the Yezo Supergroup, Hokkaido. They correlate the four subevents with the uppermost Barremian to the lowermost Aptian Stage (OAE 1a), the lower Aptian Stage (OAE 1b), the middle of the middle Albian Stage (OAE 1c) and the upper of the middle Albian Stage (OAE 1d) on the basis of ammonoid, inoceramid, and radiolarian biostratigraphy (Figure 1). This result also correlates with that of Erbacher and Thurow [21]. The timing of OAE 1a to 1d, which are detected in the Yezo Super group, appears to correlate with the ammonoid species diversity changes. The ammonoid species diversity is minimum (13 spp.) at the time of OAE 1a, the rate of extinction is 1.0, being the highest during the Cretaceous, and the rate of origination is 0.62 (Figure 3). The turnover rate at the time is -0.33 (Figure 4). It follows that all the older species are extinct but comparatively larger number of new species originated.

At the time of OAE 1b, the species diversity is at a moderate 28 species, showing a declining pattern toward OAE 1c (Figure 1). The rate of extinction is 0.70, but the rate of origination is 0.61 (Figure 3). The turnover rate is -0.12 (Figure 4) .

At the time of OAE 1c, the species diversity is 13, forming a very clear minimum (Figure 1). The rate of extinction is as large as 0.89, with a fairly large rate of origination 0.85 (Figure 3). Both two subevents, OAE 1b and 1c, show a similar characteristic in relation to the rates of extinction and origination. The turnover rate at the time of OAE 1c is -0.33, showing the lowest level during the Cretaceous (Figure 4).

At the time of OAE 1d, the species diversity is 94, being the highest in the Cretaceous (Figure 1). It is apparent in field observations that this is due to the low biostratigraphic resolution of this age in Japan. In addition to the studies of OAE 1 mentioned earlier, Maeda [24] mentioned that the benthic fossils are very rare in the Albian well laminated black mudstone beds, which extend for more than 2000 km in Japan and Sakhalin. Ammonoids are

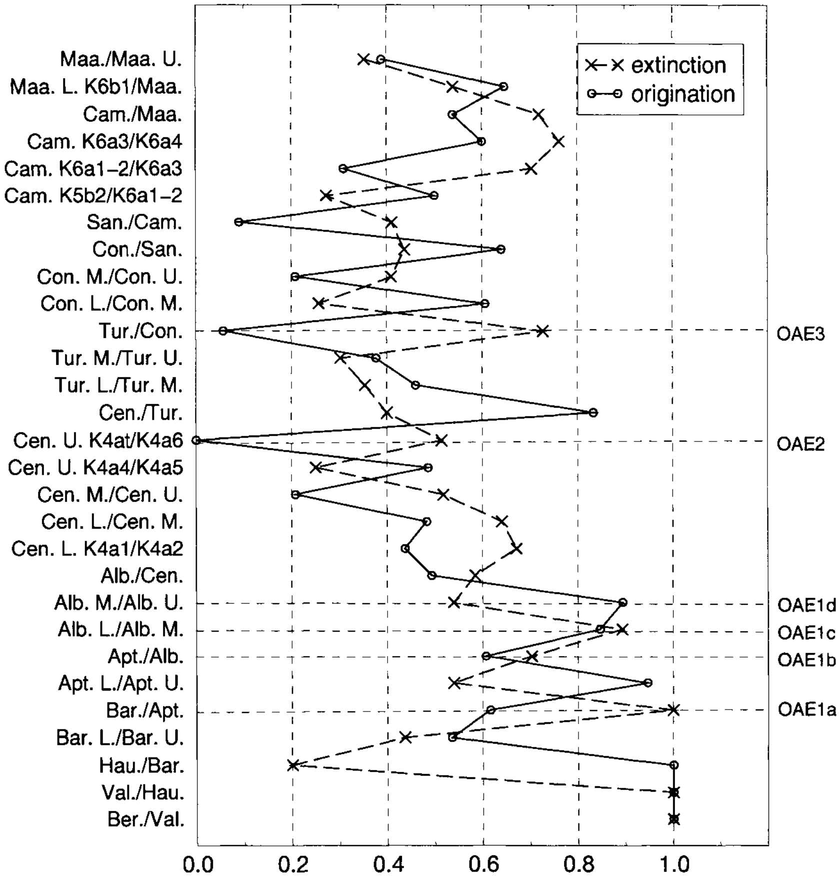

Figure 3. Rates of extinction and origination in Cretaceous ammonoid species of Japan at the boundary of each stage or substage.

also very rare in such beds. Thus, rare ammonoids in the well laminated black mudstone beds and the high occurrence of ammonoids above are totalled as the upper Albian (K3b3 in Japanese domestic substage) diversity. Black to gray mudstone prevails and various biological markers indicate dysoxic conditions through the subevents from OAE 1a to 1d in the Yezo forearc basin deposits, all of which suggest us that repeated subevents of anoxic to dysoxic conditions in addition to the significant input of terrestrial clastics prolonged dysoxic conditions [23]. Thus, ammonoid species diversity appears to have been controlled by the oceanic anoxic subevents, in which biostratigraphic resolution is sufficient for the discrimination in and after the subevents. It is also noteworthy that in OAE 1 origination of new species exists even in subevents, and recovery of the diversity occurred in the time between OAE 1a and 1b, and after the OAE 1d.

Hirano et al. [25], Hasegawa and Saito [26] and Hasegawa [27] noted the existence of OAE

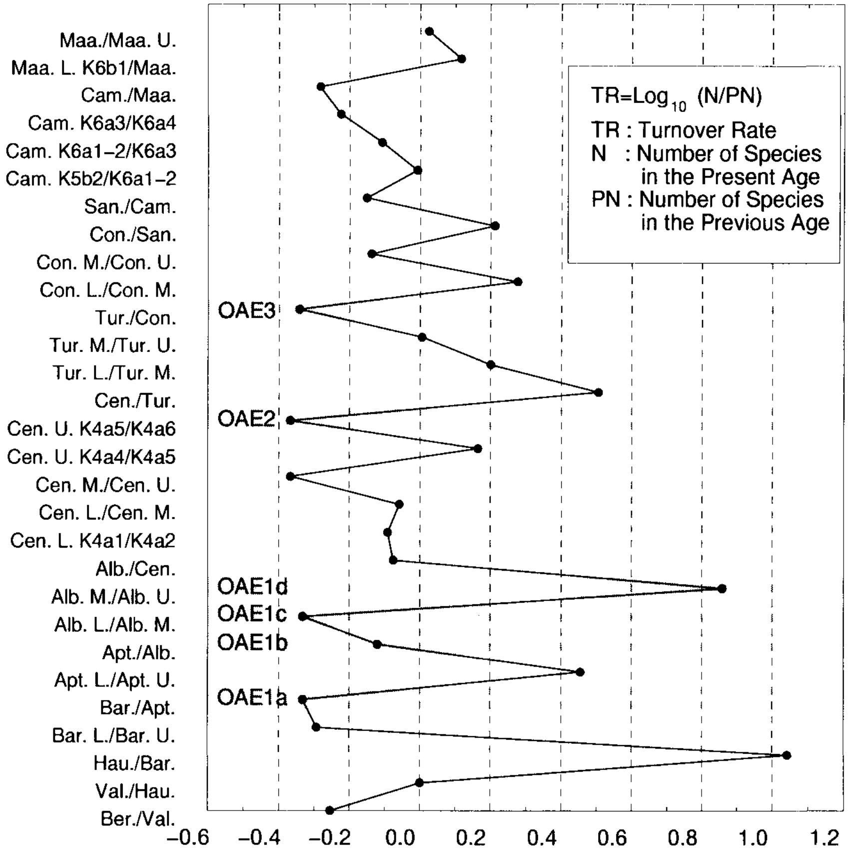

Figure 4. Turnover rate of Cretaceous ammonoid species in Japan.

2 for the Yezo Supergroup. As mentioned earlier, in the upper Cenomanian the ammonoid species diversity in Japan dropped to as low as 15, similar to OAE 1 a and 1c (Figure 1). At this time, the rate of extinction is 0.51, which is lower than those of OAE 1, but the rate of origination is zero (Figure 3) and the rate of turnover is -0.37 (Figure 4). At the time of OAE 2, the Cenomanian ammonoid *Desmoceras japonicum* was limited in its distribution, and only small populations in the nearshore shallow area survived. The bottle neck effect is easily presumed probabilistically to have worked, when the population was much decreased, and the genetic drift in survived small populations is also easily presumed genetically. Hirano [28] described gradual evolution in some characters but punctuated equilibrium in other characters of *D. japonicum*, and discussed the evolution from this species to the Turonian *Tragodesmoceroides subcostatus* in the nearshore shallow species during the time of OAE 2.

Benthic bivalves like *Inoceramus kamuy* and *Mytiloides* aff. *sackensis* and nektic ammonoid like *Puzosia orientalis* occur just above the black mudstone bed of OAE 2 [7], although fossil occurrence in the bed is very rare. The *Mytiloides labiatus* group, which is known as opportunistic species [29], occurs abruptly at the bed just above the tuff dated 87.7 $\pm$ 1.9 Ma [30]. Therefore, the condition of the basin bottom recovered 2 to 3 m.y. after the end of OAE 2, although the species group is opportunistic and may have been fairly tolerant of dysoxic conditions.

Hasegawa [27] showed radiolarian abundance (expressed by the radiolarian individuals per total grains) for the Cenomanian/Turonian stage boundary as rarely exceeding more than 40 %, and mostly below than 20 %, in *Euomphaloceras septemseriatum* zone and *Neocardioceras juddi* zone. The abundance decreases below 5 % during OAE 2, but recovered rapidly the abundance to 90 % during *Watinoceras devonense* zone. This tells us that the surface water ecosystem rapidly recovered after the end of OAE 2.

Kaiho [31] noted that the duration of the OAE 2 is 500,000 years, and a 37 % extinction ratio of calcareous benthic foraminifers, which lived between 300 m and 600 m is 10 species. Incidentally, those in the Dover section, U. K., is 7 species and the extinction rate is 58 % [32]. In contrast, the rate of extinction of planktic foraminifers is 28 to 41 %, significantly lower than benthic foraminifers. It is supported that dysoxic water did not develop in the bottom water but only in the middle and deep waters.

The Coniacian Stage is represented by transgressive sediments in Japan as described by Okada and Sakai [1]. The Coniacian black mudstone beds sometimes contain numerous specimens of *Didymotis akamatsui* exclusively, which are correlated with the European *Didymotis* event of Wood et al. [33] (see also Hirano and Takagi [34]). It is suggested that this level represents OAE 3. The ammonoid species diversity is as low as 35, showing a minimum (Figure 1). The rates of extinction and origination are 0.73 and 0.06, respectively (Figure 3). The turnover rate is - 0.34 (Figure 4).

As described and discussed above, it follows that the species diversity decreased at the times of OAEs due to the larger rate of extinction than that of origination, although there is a problem of low biostratigraphic resolution in OAE 1d. This is due to the prevailing dysoxic or anoxic conditions as suggested by the studies of the excursion of the stable carbon isotope ratios and various biological markers of Hirano and Fukuju [22] and Takahashi et al. [23]. It would be important to understand more precisely the vertical range of the dysoxic to anoxic waters in OAEs.

The rapid terminal Cretaceous extinction of marine organisms is limited to surface water organisms [31]. The ammonoid diversity gradually decreased after the Santonian (Figure 1), and inoceramid bivalves disappeared during the upper Maastrichtian before the end Cretaceous in Japan [35]. This phenomenon is same as that described by Marshall and Ward [36] for Western Europe.

5. CONCLUDING REMARKS

The diversity changes of 902 ammonoid species of the Japanese Cretaceous during the 29 substages have been described for the first time. The maximum diversity is 94 species in the upper Albian, and there are six times of minima smaller than 40 species in the Valanginian-Hauterivian, lower Aptian, middle Albian, upper Cenomanian K4a4, upper Cenomanian K4a6 and lower Coniacian. Among these minima, the four minima in the lower Aptian, middle Albian, upper Cenomanian and lower Coniacian correlate with OAE 1a, OAE 2 and OAE 3, respectively.

The occurrence of OAE 3 in Japan is firstly described in this paper based on the facts that (1) the lower Coniacian is represented by black mudstone with no benthic fossils, (2) only *Didymotis akamatsui* occurs, (3) the ammonoid diversity is low, and (4) the turnover rate of ammonoid species is minus.

The minimum diversity is not detected in the time of OAE 1d, because the duration of OAE 1d in the upper Albian K3b3 is shorter than the biostratigraphic resolution of the stage.

Thus, the species diversity of the Japanese Cretaceous ammonoids was primarily controlled by the oceanic anoxic events.

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Early Cretaceous climatic provinces in Japan and adjacent regions on the basis of fossil land plants

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Three distinct coeval floras were present in the Early Cretaceous time in Japan and its adjacent regions. They are Ryoseki-type, Tetori-type and Mixed-type floras, and each has characteristic taxa that indicate palaeoclimatic features. This paper analyzes the climatic significance of these Early Cretaceous floras on the basis of floristic taxa.

1. INTRODUCTION

Early Cretaceous plants are varied and abundant in Japan and its adjacent regions. They can be arranged as Ryoseki-type, Tetori-type and Mixed-type floras [1, 2, 3]. According to my study on the Early Cretaceous plants in Japan and its adjacent regions, the Ryoseki-type floras grew in subtropical and tropical environments with annual dry season; the Tetori-type floras grew in temperate regions; and the Mixed-type floras, with older angiosperms, are represented by either a predominance of the Ryoseki-type taxa or by a predominance of the Tetori-type taxa. The origin of angiosperms has long been the subject of much debate. During the past decade, many Early Cretaceous (or Late Jurassic) angiosperms were recorded in association with the Mixed-type floras (e.g. Ohana and Kimura [4]). However, these angiosperms have not been found in the properly known coeval Ryoseki-type and Tetori-type floras in Japan and adjacent regions.

2. THE RYOSEKI-TYPE FLORAS

In East and Southeast Asia, the Ryoseki-type floras are extensively distributed as shown in Figure 1. In Japan the Ryoseki-type floras are known from the scattered areas along the Outer Zone (Pacific side). Most plant-bearing formations are of marine origin, so that their geological ages are determined with considerable accuracy. The most representative taxa of the Ryoseki-type floras are those belonging to cheirolepidiaceous conifers (e.g. *Frenelopsis*), which have

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a very thick cuticle and remarkably sunken stomata.

Furthermore, typical subtropical-tropical taxa (genera) known in the Ryoseki-type floras are shown below (* = typical Ryoseki-type genera; ** = systematic position unknown):

Bryophyta; *Thallites*
Lycopodiales; *Lycopodites*
Sphenopsida; *Annulariopsis, Neocalamites, Equisetites*
Ferns or fern-like plants; *Nathorstia, Todites, Klukia, Gleichenites**, *Coniopteris* (very rare), *Eboracia* (very rare), *Cyathocaulis**, *Yezopteris**, *Weichselia**, *Matonidium**, *Onychiopsis, Acrostichopteris, Adiantopteris, Sphenopteris***, *Cladophlebis***
Seed ferns; *Sagenopteris, Stenopteris, Ptilozamites***, *Pachypteris**

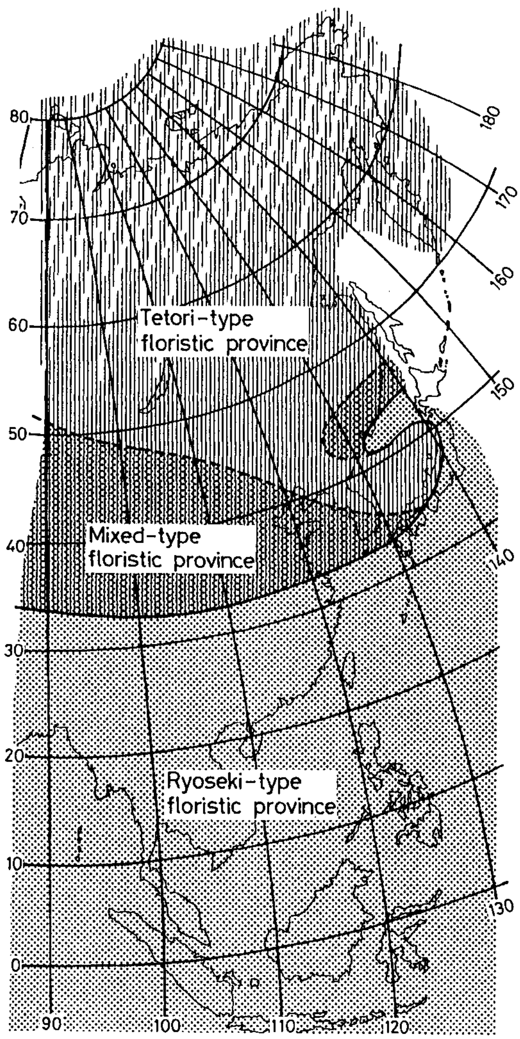

Figure 1. Distribution of three types of floras of Late Jurassic - Early Cretaceous time, in Eastern Eurasia and Southeast Asia. Detailed localities of fossil plants were already shown by Ohana and Kimura [4]. The vegetation of northern Siberia is essentially of the Tetori-type, but in this region, bennettitalean and cycadalean taxa are less in number and not abundant.

Bennettitaleans; *Ptilophyllum*, Pterophyllum, Nipponoptilophyllum*, Zamites*,Otozamites, Williamsonia*, Weltrichia*, Bucklandia**

Cycadaleans; *Nilssonia, Sanchucycas*

Conifers; *Parasequoia, Pagiophyllum, Elatocladus, Nageiopsis, Podocarpus (Nageia)*, Frenelopsis*, Pseudofrenelopsis*, Brachyphyllum*, Cyparissidium, Cupressinocladus*, Mesembrioxylon**

Incertae Sedia; *Taeniopteris**, Geonoma**, Phillites***

In general, the Ryoseki-type taxa are characterized by :

1) Pinnules of ferns are small-sized (e.g. *Weichselia*: undetermined form-genera such as *Sphenopteris* and *Cladophlebis*), coriaceous and reflexed at margins.
2) Pinnae of bennettitaleans and cycadaleans are thick and coriaceous (e.g. *Nilssonia* sp. cf. *N. schaumburgensis* (Dunker) Nathorst [5]).
3) Conifers with needle-like leaves (e.g. *Parasequoia*) are rather rare, and conifers with scale-like leaves are common (e.g. *Frenelopsis*).
4) Cuticles of bennettitaleans, cycadaleans and conifers are commonly thick.
5) No remarkable annual rings in wood remains were observed.

These features might show that the Ryoseki-type floras grew in the subtropical-tropical regions with annual dry season.

3. THE TETORI-TYPE FLORAS

In East Asia, the Tetori-type floras are distributed in northern territories as shown in Figure 1. In Japan this flora is derived from the Tetori Supergroup which is partly of marine origin, and this supergroup is located extensively in the Inner Zone (Japan Sea side) of Central japan. Sediments of the Tetori Supergroup include occasional coal seams, while no seams are found from the sediments yielding the Ryoseki-type plants.

Fossil plants are varied and abundant in the black or greyish shales. The known taxa (genera) are shown below (* = typical Tetori-type genera; ** = systematic position unknown):

Bryophyta; *Thallites*

Sphenopsida; *Equisetites*

Ferns and fern-like plants; *Osmundopsis, Todites, Raphaelia*, Gleichenites, Coniopteris*, Dicksonia*, Eboracia*, Birisia (=Acanthopteris)*, Hausmannia, Asplenium, Arctopteris*, Endoa* (water fern), *Onychiopsis, Adiantopteris, Sphenopteris**, Cladophlebis***

Seed ferns; *Ctenozamites, Sagenopteris*

Bennettitaleans; *Otozamites, Anomozamites, Pterophyllum, Pseudocycas, Neozamites*, Dictyozamites**

Cycadaleans; *Ctenis*, Tetoria*, Nilssonia* (including *Nilssoniocladus*)

Cycadophyte with unknown affinity; *Cycadites*

Ginkgoaleans; *Ginkgoidium*, Pseudotorellia*, Eretomophyllum*, Ginkgoites*, Baiera*, Sphenobaiera**

Czekanowskialeans; *Czekanowskia*, Phoenicopsis*, Arctobaiera*, Leptostrobus**

Figure 2. A supposed vegetational view of the Early Cretaceous Ryoseki-type flora (drawn by T. Kimura).
1-2: small-sized horse-tails, *Neocalamites* (1) and *Annulariopsis* (2). 3: a gleicheniaceous fern. 4-5: matoniaceous ferns, *Weichselia* (4) [9] and *Nathorstia* (5). 6: dicksoniaceous ferns, *Onychiopsis* of the Ryoseki-type. 7: an *Adiantopteris*-resembling extant fern, *Adiantum*. 8-10: bennettitaleans, *Ptilophyllum* (8), *Zamites* (9) and *Williamsoniella* (10). 11: a cycadalean with slender stems, *Nilssonia*. 12: araucariaceous or podocarpaceous conifers, *Nageiopsis*. 13: conifer with scaly leaves, *Cyparissidium* or *Frenelopsis*.

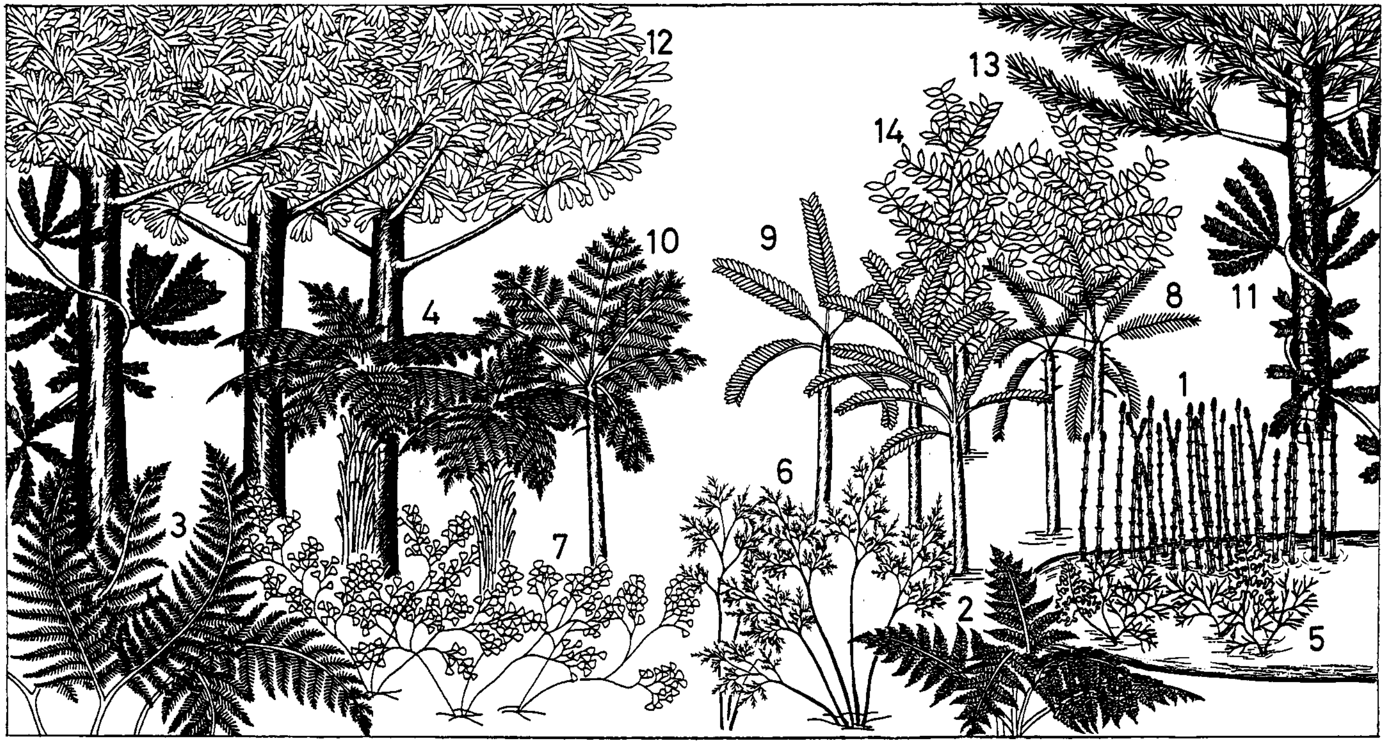

Figure 3. A supposed vegetational view of the Early Cretaceous Tetori-type floras (drawn by T. Kimura).
1: small-sized horse-tails, *Equisetites*. 2: an osmundaceous fern, *Todites*. 3: a gleicheniaceous fern, *Gleichenites*. 4: a dicksoniaceous tree fern, *Coniopteris*. 5: parkeriaceous water ferns, resembling extant *Ceratopteris*. 6: a dicksoniaceous fern, *Onychiopsis* of the Tetori-type. 7: *Adiantopteris* of the Tetori-type, externally resembling the extant *Adiantum*. 8-9: bennettiataleans, *Dictyozamites* (8) and *Otozamites* (9). 10-11: cycadaleans, *Tetoria* (10) and *Nilssonia* (11). 12: ginkgoalean trees. 13: a czekanowskialean tree. 14: broad leaved conifer tree, *Podozamites* with unknown affinity.

Conifers; *Pityophyllum, Elatocladus, Podozamites*, Lindleycladus*, Protodamarra*
Systematic position unknown; *Tetoriophyllum**, Jucutopteris**, Butefia* (?)**, *Taeniopteris**, Aphlebia**, Carpolithus**, Xenoxylon***

The most representative taxa of the Tetori-type floras are ginkgoaleans. Generally, they occur abundant locally. They are not found from the Ryoseki-type floras, except for a few leaf fragments from the Early Cretaceous Jabalpur plant beds in India (e.g. Bose and Sukh-Dev [6]). These ginkgoalean leaves are deciduous [7]. According to Sogame [8], extant ginkgoaleans are planted in subtropical and temperate regions. I am of the opinion that the Cretaceous ginkgoaleans are also subtropical-temperate habitants like those of the extant *Ginkgo biloba*. Because the extant and fossil ginkgoaleans (e.g. *Ginkgoites*) show similar growing habit and deciduous leaves.

In general, the Tetori-type taxa are characterized by:

1) Pinnules of ferns are rather large-sized with some exceptions and thin in texture.
2) Serrate or spinose pinna margins are present in bennettitaleans (e.g. *Neozamites*) and cycadaleans (e.g. *Ctenis* and *Nilssonia*).
3) *Nilssonia* leaves are generally deciduous and short-shoots of czekanowskialeans are also deciduous. The deciduous habit indicates that these taxa are temperate habitants. Similar deciduous habit is seen in the extant angiosperm leaves with marginal serration.
4) Annual rings are conspicuous in wood remains (e.g. *Xenoxylon*).
5) Coal-seams are occasionally present in the Tetori sedimentary basins in Japan, but are fairly developed in Northeast China and Siberia.

The facts mentioned above might show that the Tetori-type floras grew mainly in the temperate regions in Early Cretaceous time.

Figures 2 and 3 show supposed vegetational views of the Early Cretaceous Ryoseki-type and Tetori-type floras, respectively.

The difference between the Tetori-type and the Ryoseki-type floras is notable in floristic character. It is likely that in Latest Jurassic-Early Cretaceous time, both types of floras were flourished in different floristic regions.

4. THE MIXED-TYPE FLORAS

This type of floras are known between the Tetori-type and the Ryoseki-type floristic realms (Figure 1). The Early Cretaceous angiosperms were found in the Mixed-type floras. These angiosperms are mostly represented by the small-sized, irregular leaf-form and venation, and by having entire leaf margins.

It is likely that these angiosperms flourished in the subtropical or warm-temperate regions corresponding to the Mixed-type floristic realm.

5. CONCLUDING REMARKS

In Latest Jurassic-Early Cretaceous time, two distinct floras, the Ryoseki-type and the Tetori-type, flourished. On the basis of known taxa, the Ryoseki-tupe floras flourished in the subtropical - tropical regions with annual dry season. This type of floras were distributed in the Outer Zone of Japan, Southern Primorye, South China and Southeast Asia.

The Tetori-type floras flourished in the temperate regions. This type of floras were distributed in the Inner Zone of Japan, North and Northeast China and Siberia.

Between both floristic realms, the Mixed-type floras were present with older angiosperms (Latest Jurassic-Early Cretaceous) indicating subtropical or warm-temperate climate.

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Cretaceous active margin in the eastern Eurasia continent affected by the Southern Pacific superplume

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Influences of the Southern Pacific superplume upon the Cretaceous eastern Eurasia are investigated by the mantle flow model. The plume impacts on the sublithosphere mantle flow and stimulates the Kula and Pacific plate motions. The perturbations of the mantle flow and the plate motions activated the upwelling flow of the mantle through the subducted Kula-Pacific ridge gap beneath the Eurasia margin; transferring heat to the lower crust or uppermost mantle and creating large magma chambers have been promoted by adiabatical heat convection due to the mantle upwelling flow. The tectonic events of the continental plate also occurred especially in the fore-arc region. About 5000 km of the eastern margin in the Eurasia continent was positively affected by the superplume eruptions; the active marginal plutonism as well as rhyolitic to andesitic volcanism occurred together with hard tectonic events and they migrated northeastward along the continental margin.

1. INTRODUCTION

Cretaceous magmatism was conspicuously active in the eastern margin of the Eurasia continent. Large volumes of granitic rocks and their volcanic equivalents crop out along the 8000 km continental margin from the southwestern part of Guangxi in South China to the Chukot peninsular in the Far East of Russia [1, 2]. The migration model of magmatism explaining the ridge subducton was first proposed by Kinoshita and Itô [1, 3], due to the northeastward younging of the granite age along the Eurasia continental margin. The subducted active ridge was the Kula-Pacific ridge which opened the slab window under the continent and migrated northeastward along the continental margin [4, 5]. The conspicuous magmatic activity requires further incentives to affect on the margin, although the ridge subduction model is essentially related to active magmatism. The Southern Pacific superplume in the mid-Cretaceous may have had a strong positive effect upon the sublithosphere mantle flow and the ocean plate motions. Thus, the overlaid continental plate was seriously affected in the arc and fore-arc region.

The superplume as the long-term heat cycle of the earth caused the latest pulse in the mid-Cretaceous (e.g., [6]). The superplume was very active and extensive in the southern Pacific Ocean (ca. 5000 km across). The plume-related events have been reported from the Circum-Pacific rim [6, 7, 8]. Sedimentary basins related to the plume activity were also presented in detail in Cretaceous East Asia [9]. Below I discuss the effects of the Southern Pacific superplume on the Cretaceous magmatisms and the other tectonic events in the eastern margin of the Eurasia continent in the context of a mantle flow model and plate tectonics.

2. SUPERPLUME-EFFECTS ON THE EASTERN EURASIA MARGIN

The mantle circulation is modeled with reference to Maruyama [7] as shown in Figure 1. In the ordinary stage, the upper mantle convects from the divergent (ridge) to the convergent (trench) zones (solid arrows in the control surface in the figure). It is induced by the oceanic plate motion in the upper part of the upper mantle. The flow that descends with the slab in the trench zone turns back in the lower part of the upper mantle to the base of the ridge zone. The descended slab accumulates at the base of the upper mantle as a cold megalith. The megalith is almost stagnant, however it grows as large as the slab descends and creates additional flow to turn back to the base of the ridge zone. Thus, the volume of returning flow to the ridge zone is continuously kept to be the ordinary circulation volume. At some point, either as a result of excess mass accumulation, or as a result of the ascent of a hot superplume flow from the core mantle boundary, a part of accumulated megalith descends into the lower mantle. When this happens the normal mantle circulation is bypassed as much as the volume of descending megalith, adding hot lower mantle material to the base of the lithosphere. The superplume just breaks out.

An assumption of the mantle flow model is that the source volume $2\pi K$ of the superplume is added to the uniform flow (velocity) U moving with the oceanic lithosphere toward the trench

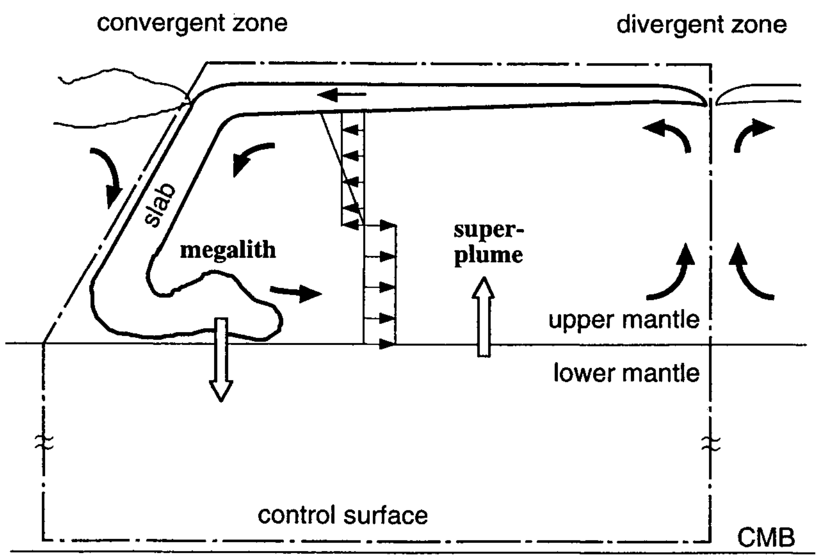

Figure 1. Mantle flow model circulating in a two-layer system with upper and lower mantle convectively isolated. Upper mantle convection normally occurs as indicated by solid arrows in the control surface. Episodically, bypass flow happens (hollow arrows). This coincides with breakthrough of the superplume

zone. When the potential flow is assumed in this case, the complex potential W(z) of the flow superposed with the source flow upon the uniform flow is W(z) = Uz + K ln z. The assumption of potential flow may be made for the flow field apart from the source center where the flow is uniform in the vertical plane. The flow pattern of the above equation is easily drawn by Rankin's method (Figure 2), placing the source at the center of the superswell [10] in the Southern Pacific. The flow conditions are taken as follows. The strength of the source is assumed to be as large as the Ontong-Java plateau, in which the formation rate of the plateau is estimated to be 14 km^3/yr [11] and 17 km^3/yr [12]. In this case, the total source volumes are calculated to be 70 and 85 km^3/yr, respectively, with the assumption that the plateau has been produced by 20 % partial melt of the flow. In the convergent zone shown in

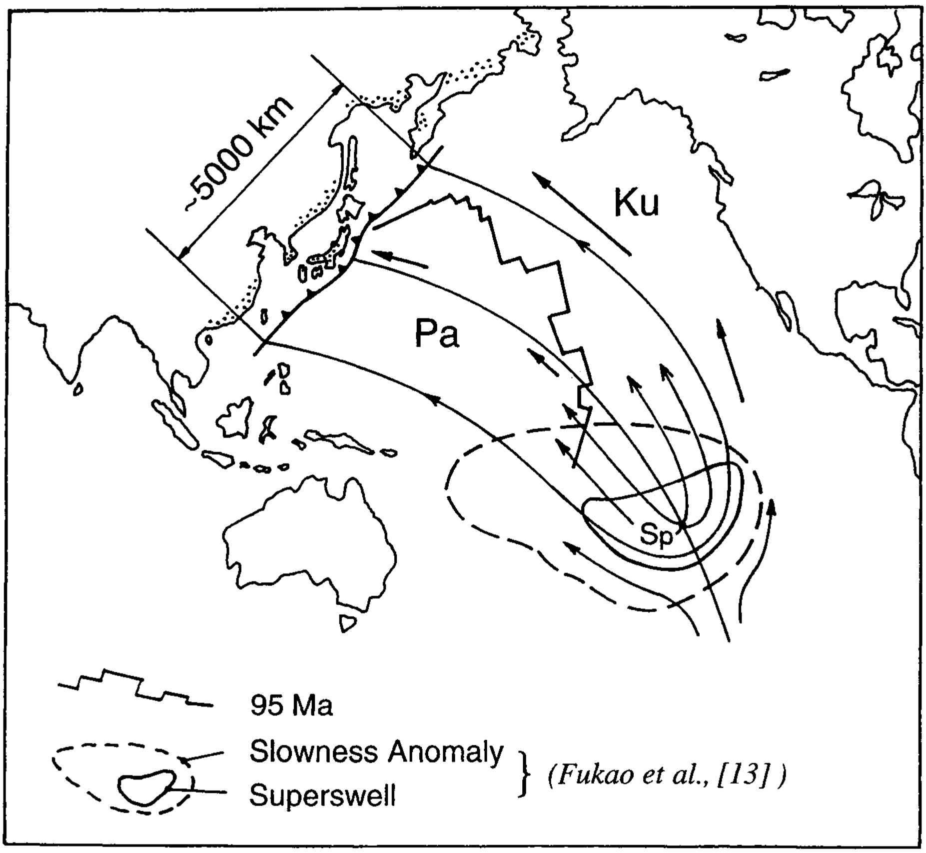

Figure 2. Flow pattern superposing a superplume source upon uniform upper mantle flow. The source center is put at the center of the superswell. The shallower mantle (Figure 1) uniformly flows northwestward, which is almost consistent with the Kula and Pacific plate motions. Ku: Kula plate, Pa: Pacific plate and Sp: superplume, dotted areas: magmatic belt in the east Asia margin.

Figure 1, taking 5 cm/yr and 300 km as the average velocity and the depth of the mantle flow, respectively, the mantle flow widths in the cases of source volumes of 70 and 85 km^3/yr are estimated to be 4700 and 5600 km, respectively. Thus, expansion of the flow causes about 5000 km of the eastern Eurasia continental margin to be affected by the superplume. The dotted areas in the eastern Asia margin in Figure 2 show the magmatic belts.

In the Southern Pacific in Figure 2, a topographic anomaly (superswell) about 4800x2300 km in extension and the slowness anomaly of seismic velocity about 8800x5000 km at 2100 km depth under the superswell are closely related and reflect the surface geology and the thermal structure formed by the superplume [13]. The border of the flow is the dividing streamline that separates the flow into two parts, the inner superplume flow and the outer uniform flow. Because the transient time in the high viscosity flow is very short, the flow pattern drawn in the figure arises soon after the source flow is added. When an active ridge like the Kula-Pacific ridge was located in the plume flow (as shown by 95 Ma ridge in the figure), upwelling through the subducted ridge gap beneath the continental margin and the rate of motion of both plates may have been strongly affected by the superplume pulse. Not only wide plutonism but also the active acidic to intermediate volcanisms occurred during 130 Ma in the western Fujian, South China to 70 Ma in the central Sikhote Alin, Russia.

Tectonic events in Japan area were the great hiatus (125-90 Ma) in accretion in the Shimanto belt and the formation of the Median Tectonic Line (about 100 Ma). The hiatus locates at the Butsudo Tectonic Line between the Sanbosan and the Northern Shimanto belts [14]. They were occurred by the perturbation due to the motion of the Kula or Pacific plate in the superplume eruptions.

3. CONCLUDING REMARKS

The Southern Pacific superplume was a wide and active volcanism which generated very large plateaus and sea mounts in the very short period (for a few million years) in the Polynesia oceanic region. In the eastern margin of the Eurasia continent, the strong arc-magmatism also occurred together with the active tectonic events. I think both had the peak activity in the mid-Cretaceous and were not independent, although they were distant (more than 10000 km) each other. That is, the Eurasia marginal magmatism was positively activated by deeper pulsation of the Earth, superplume eruption and closely controlled by shallower tectonic structure, ridge subduction system.

The chief characteristics of the Cretaceous eastern Eurasia arc-magmatism was of course strong and migrating activity, which is indicated by extensive out-crops of granitic and volcanic rocks and the northeastward-younging of the granite age trend along the continental margin. The Southern Pacific superplume stimulated it into more vigorous activity by enhancing plate motioms and asthenosphere flow. Its effect upon the Eurasia continent was about 5000 km along the margin.

However, clear evidences of the superplume activity are not so many to discuss how strong the effect was on ridge subduction system with slab window. So far from discussing it, superplume itself is yet left to be a difficult problem to drive to idea's end. It is certain that the superplume and ridge subduction are big events seriously affecting on the Cretaceous environment. This is the reason why proposition of assumption or speculation related to plume effects on the Eurasia margin was done by the mantle convection model, although verification of this assumption based on geoscientific data is left in next paper. It is a thoughtprovoking topic of the day to derive assumption for superplume episode by various methods and to check it by many evidences.

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Comparison of genesis and tectonic setting of Jurassic and Cretaceous high-pressure metamorphic belts in the circum-Pacific regions

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A large number of serpentinite seamounts in the western Pacific Izu-Ogasawara-Mariana forearcs indicate the serpentinite diapirism under the tensional stress field in the forearc regions. The presence of blueschist-facies rocks in the serpentinite seamounts provides evidence that the serpentinites generated from the hanging-wall peridotites have penetrated the high-pressure metamorphic region at depth in the accretionary prism. The high-pressure metamorphic terranes in the circum-Pacific regions, the Franciscan Complex, California, and the Kamuikotan metamorphic belt, Japan, are accompanied with serpentinite melange with tectonic blocks of deeper high-grade metamorphic rocks, whereas the Sanbagawa metamorphic belt, Japan, does not contain any serpentinite melange. It is highly probable that the former were formed under tensional stress regime, and the latter under compressional one.

1. INTRODUCTION

Jurassic to Cretaceous high-pressure/low-temperature type metamorphic terranes are widely extended in the circum-Pacific regions. Studies of these terranes in the light of plate tectonic theory have revealed that they represent ancient subduction complexes. In the Izu-Ogasawara (=Bonin)-Mariana forearc regions, a large number of seamounts that consist of serpentinized peridotites are widely distributed [1]. During ODP Leg 125, high-pressure blueschist-facies clasts were found from Conical Seamount, one of the seamounts in the Mariana forearc. It suggests that a high-pressure metamorphism is ongoing within the active subduction zone [2, 3]. The discovery of blueschist-facies rocks in the modern forearc region makes it possible to discuss the genesis of high-pressure metamorphism and related tectonics in terms of comparison between on-land high-pressure metamorphic terranes and modern subduction system. In the high-pressure metamorphic terranes, serpentinites are intimately associated with *in situ* crystalline schists derived from pelagic sediments and volcanics. The modes of occurrence of serpentinites in these terranes provide clue to restoring tectonic environments

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around high-pressure metamorphism and are important for understanding how the mantle materials were taken into the sediments of accretionary prism. In this paper, I will summarize the tectonic framework on high-pressure metamorphism and mode of emplacement of serpentinites in the Izu-Ogasawara-Mariana arc-trench system, and compare them with the Franciscan Complex, California, and Kamuikotan metamorphic belt, Japan; the tectonic frameworks in and around these on-land metamorphic terranes are closely akin to those of modern arc-trench systems. On the other hand, the Sanbagawa metamorphic belt, Japan, appears to have a quite different tectonic framework from these regions. I will also discuss the tectonic framework of the Sanbagawa metamorphic belt and consider an essential factor controlling the tectonic environments during high-pressure metamorphism.

(A)

(B)

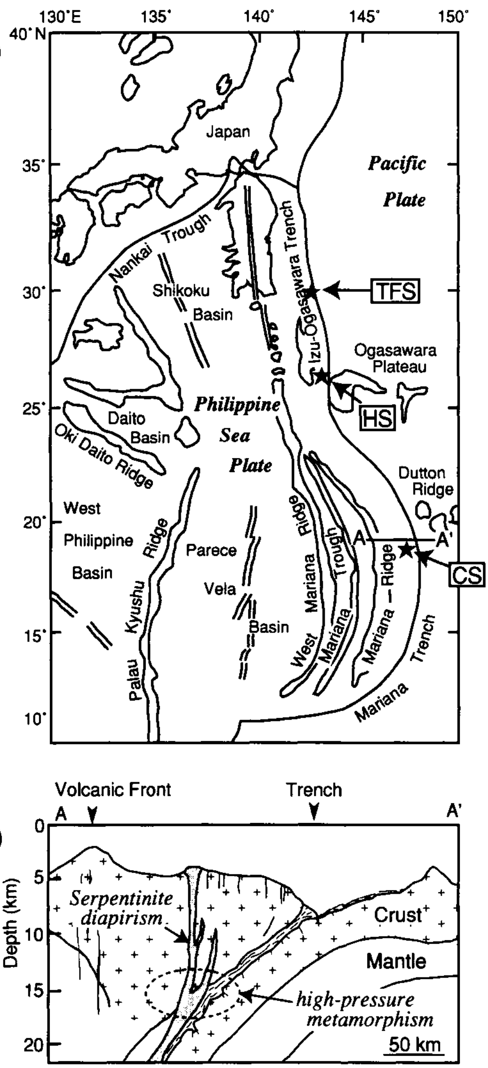

Figure 1. (A) Regional map of the western Pacific Ocean. TFS: Torishima Forearc Seamount, HS: Hahajima Seamount, CS: Conical Seamount. (B) Schematic cross section showing the tectonic framework of the Mariana arc-trench system (based on the data from Hussong and Uyeda [39] and Maekawa et al. [7]).

2. TECTONIC FRAMEWORK IN THE MODERN IZU-OGASAWARA-MARIANA ARC-TRENCH SYSTEM

2.1. Serpentinite diapirism

In the Izu-Ogasawara-Mariana forearc regions, a chain of seamounts extends for more than 2500 km along the trench axis [1, 4] (Figure 1). The seamounts are commonly dome-like in shape, and are up to 30 km in diameter with up to 2 km of relief. These seamounts are composed mainly of serpentinized peridotites. Alvin submersible dives on the flank of Conical Seamount in the Mariana forearc revealed that serpentinite blocks of varying sizes are scattered in a fractured and crushed serpentinite mud matrix on the surface of the seamounts, and carbonate and silicate chimneys are formed on the southwest side of the summit [5]. Drilled samples of serpentinites from ODP Leg 125 are highly sheared, and show "block-in-matrix" fabrics, which are typical in on-land serpentinite melanges. The fluids actively seeping from the chimneys and pore fluids from the summit drill samples are characterized by high pH, more than one-half deficiency of Cl and Br relative to seawater, and enrichment in CH_4, SiO_2, SO_4, and H_2S [5, 6]. The geochemical analysis suggests that the fluids have been generated during serpentinization at the top of the downgoing slab, 30 km below the seafloor. When the oceanic plate starts to subduct below the other plate, the descending oceanic slab must encounter the hanging-wall mantle peridotites at depth. As hydrated pelagic sediments at the top of the subducting slab supply water to peridotites, the peridotites react with water to form low-density serpentinites. The forearc regions were intensively destroyed by many normal faults caused by the tensional stress in the west of the trenches in the Izu-Ogasawara-Mariana forearcs. Voluminous and low-density serpentinites generated just above the subducting slab have uplifted along the normal faults to form a huge chain of seamounts on the ocean floor.

2.2. Blueschist-facies metamorphism

Recent drilling and dredging have revealed that the forearc seamounts commonly contain metamorphosed mafic rocks of blueschist facies in the Izu-Ogasawara-Mariana regions. Drill cores from Hole 778A at Conical Seamount during ODP Leg 125 contain 2- to 5-cm fragments of metamorphic rocks with low-grade blueschist-facies minerals, such as lawsonite, pumpellyite, sodic pyroxene, blue amphibole, and aragonite. Petrological study indicates that the approximate metamorphic conditions of these rocks are 150-250°C and 5-6 kb, and are situated at the lowest-pressure and lowest-temperature field in the blueschist-facies metamorphism [2, 3, 7]. Dredged metabasite samples from the two serpentinite seamounts in the Izu-Ogasawara forearc also contain low-grade high-pressure metamorphic minerals (Figure 1). Aegirine-augite and pumpellyite were found in dredged samples from Torishima Forearc Seamount during KH87-3 cruise, Ocean Research Institute, University of Tokyo [8], and the diagnostic assemblage of pumpellyite-actinolite facies; pumpellyite + actinolite + magnesioriebeckite + quartz + albite were found from Hahajima Seamount during KH-84 cruise [9]. The high-pressure metamorphic rocks appear to be common in the Izu-Ogasawara-Mariana forearc seamounts.

Serpentinite diapirism, which comes up from depth to the seafloor, must have trapped crust materials situated within the pathway, and entrained them in a fluidized melange rising to the seafloor. Thus, it is reasonable to assume that the blueschist-facies clasts from Hole 778A formed at about 16 to 20 km below the seafloor after initial subduction and were then entrained by uprising fluidized serpentinite materials. The presence of blueschist-facies rocks in the forearc seamounts confirms that a blueschist-facies metamorphism actually took place beneath the forearc in the Izu-Ogasawara-Mariana arc-trench system, and provides direct evidence that blueschist-facies metamorphism developed below the forearcs.

3. ON-LAND HIGH-PRESSURE METAMORPHIC TERRANES

3.1. Franciscan Complex

The Franciscan Complex of the California Coast Ranges is well known as an ancient subduction complex metamorphosed under high-pressure blueschist-facies conditions during

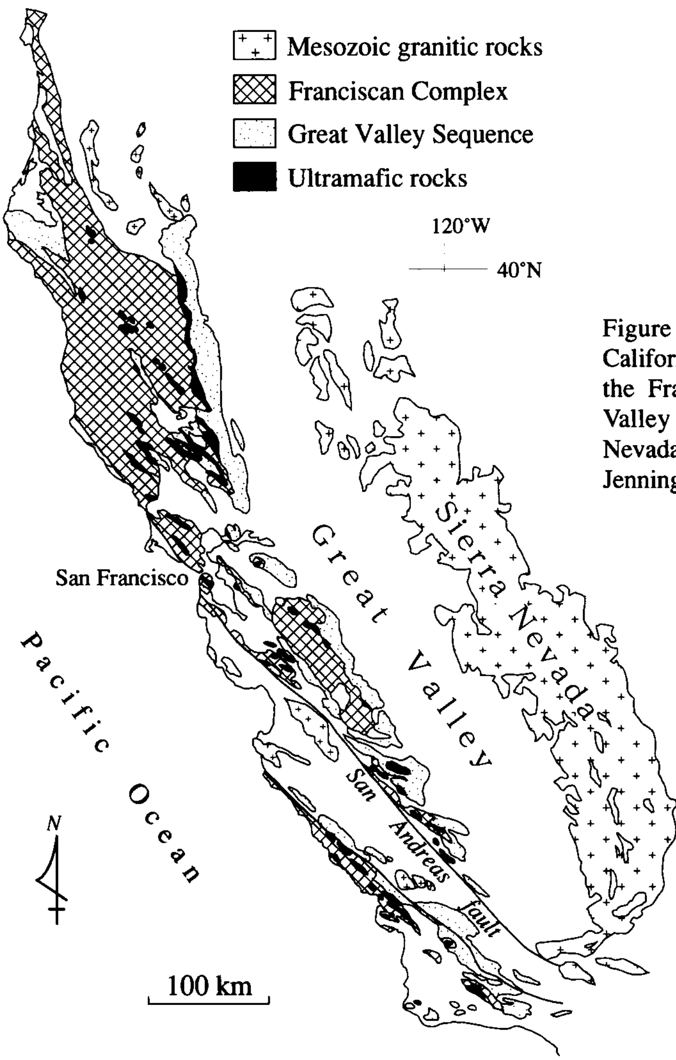

Figure 2. Geologic outline of California, showing locations of the Franciscan Complex, Great Valley Sequence, and Sierra Nevada batholith (adopted from Jennings [40]).

Jurassic to Cretaceous time. The metamorphic protoliths consist of graywacke sandstone and shale with subordinate chert and basaltic volcanics, and minor limestone. Although the original pattern of continuous belts parallel to the edge of the continent has been disturbed by 300 to 600 km of Tertiary strike-slip displacement along the San Andreas Fault system [10], approximate tectonic framework from west to east during metamorphism seems to have been well preserved (Figure 2). From west to east, the Franciscan Complex, Great Valley Sequence, and Sierra Nevada batholith are exposed to form sub-parallel zonal arrangement of once developed arc-trench system [11,12]. The Great Valley Sequence represents forearc basin sediments consisting of unmetamorphosed alternating beds of sandstone and mudstone deposited coeval with the Franciscan metamorphism. The sequence is exposed in the eastern side of northern Franciscan and in the surrounding areas around southern Franciscan. About 200 km east to the Franciscan Complex, Sierra Nevada batholith is widely exposed. It represents the province of arc magmatism contemporaneous with the Franciscan metamorphism.

The Franciscan Complex is considered to have developed under circumstances similar to the Izu-Ogasawara-Mariana subduction zone, which is accompanied by serpentinite diapirism in the forearc regions. In such subduction zones, the serpentinite protrusion pierces the region of high-pressure metamorphism below forearcs within subduction zone. In the Franciscan Complex, serpentinite masses of varying sizes are commonly recognized; some contain tectonic blocks of high-pressure metamorphic rocks, of which grade is higher than that of neighboring rocks [13]. These serpentinites may represent diapiric serpentinite melange generated from the hanging-wall peridotites just above the subducting slab and penetrating high-pressure metamorphic regions. The Great Valley Sequence comprises the serpentinite breccias of sedimentary origin at its base [14, 15]. Lockwood [16] first proposed an idea of serpentinite protrusion in the forearc regions to explain the occurrence of serpentinite breccias. Carlson [17] suggested that the breccias are formed under the similar tectonic environment to the serpentinite diapirism in the Izu-Ogasawara-Mariana forearcs. Cowan and Page [18], Moore et al. [19] and Platt et al. [20] described low-grade blueschist-facies metaconglomerates in the Franciscan Complex. These pebbles of metaconglomerate may have experienced recycling by which high-grade blueschists transported by serpentinite diapirism to sea floor were deposited at the trench, then were squeezed down and again metamorphosed under low-grade blueschist-facies conditions.

3.2. Kamuikotan metamorphic belt

The Kamuikotan belt runs north to south along the axial zone of Hokkaido, northern Japan. It extends for more than 300 km with a maximum width of 30 km (Figure 3). The Kamuikotan rocks consist of mainly lawsonite-glaucophane type metamorphic rocks, of which protoliths are pelites, chert, volcanics, and ultramafic rocks with subordinate amounts of psammites and limestone. Most of structural, petrologic, and chemical data suggest that the belt is a trace of the ancient consuming plate boundary with westward-dipping subduction [21]. A similar tectonic framework to the Franciscan is recognized in and around the Kamuikotan belt. Cretaceous granitic rocks and Lower Cretaceous volcanics are exposed in the regions 100 to 200 km west of the Kamuikotan belt. These rocks may have been the products of arc magmatism related to the subduction, which has given rise to the Kamuikotan metamorphism. The Cretaceous Yezo Supergroup unconformably overlies or locally is in fault contact with the Kamuikotan metamorphic rocks. The Yezo Supergroup consists of unmetamorphosed alternating beds of sandstone and shale coeval with the Kamuikotan metamorphism, and represents forearc basin deposits [22]. Similar to the Great Valley Sequence, the Yezo Supergroup contains detrital serpentinites and pebbles of crystalline schists [23-25]. Large amounts of serpentinite are found in the Kamuikotan belt. The Shirikoma-dake and Yubari-dake areas are underlain by large serpentinite melanges with various kinds of blocks, such as blueschist, amphibolite, metasediments, and metabasites [26, 27]. The N-S trending serpentinite melange in the central part of the Biei area includes numerous 10-m to 1-km blocks of amphibolite and blueschist. The mineral parageneses and degrees of recrystallization of these blocks suggest that they have experienced two stages of metamorphism; one is of higher-grade blueschist or epidote amphibolite facies and occurred before the serpentinite protrusion at depth, and one is lower grade blueschist-facies metamorphism which is prevailed throughout the Biei area during and/or after the serpentinite protrusion [28]. The serpentinite melange unit in the Biei area is now

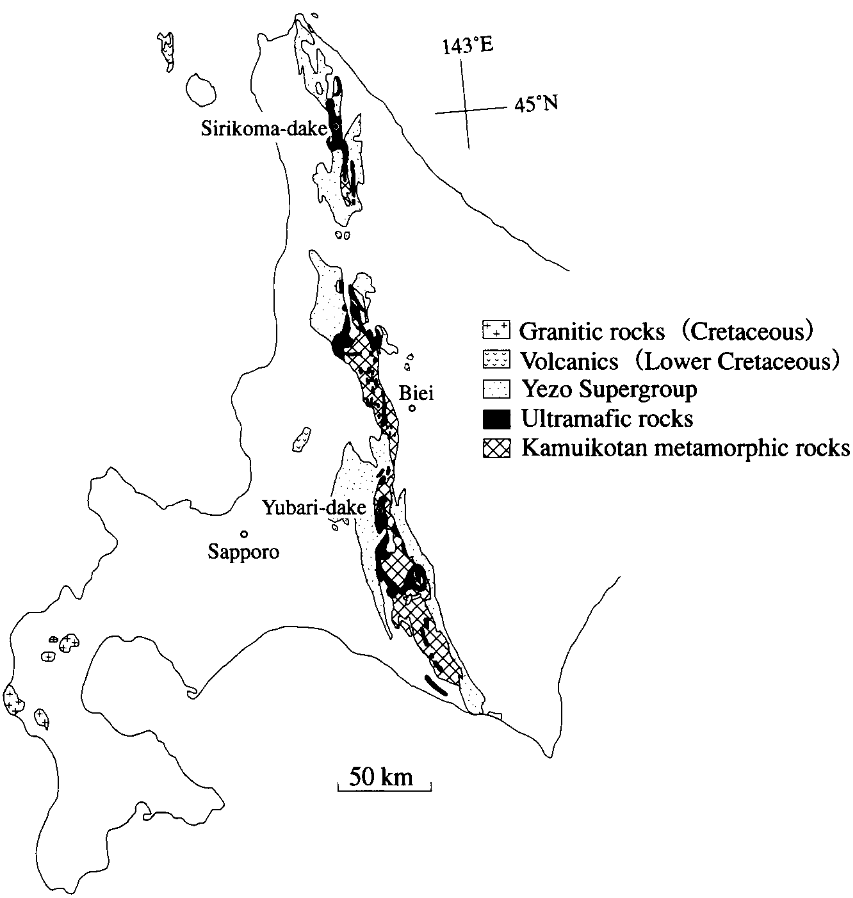

Figure 3. Geologic outline of Hokkaido, showing locations of the Kamuikotan metamorphic belt, Yezo Supergroup, and Cretaceous igneous rocks (based on the data from Takahashi et al. [41]).

bordered with adjacent units by later-stage faults, and does not seem to retain original relations with the adjacent units. It is, however, considered that serpentinite diapirism coming up from depth must have trapped higher-grade metamorphic rocks situated within the pathway, and entrained them into the lower-grade metamorphic regions where the in situ Kamuikotan metamorphic rocks have been produced. The trapped metamorphic rocks in the serpentinite melange were again metamorphosed under lower-grade blueschist-facies conditions.

3.3. Sanbagawa metamorphic belt

The Sanbagawa belt runs east to west for about 800 km throughout southwest Japan (Figure 4). The belt is one of the most extensively studied high-pressure metamorphic complexes in Japan. Petrological studies of the Sanbagawa belt have shown that the metamorphic grade can be divided into four zones based on the appearance of index minerals in pelitic schists. In increasing order of grade, these zones are the chlorite zone, the garnet zone, the albite-biotite zone and the oligoclase-biotite zone [29]. The Sanbagawa belt is bounded to the north by a major strike-slip fault, the Median Tectonic line, which separates it from the Cretaceous low-pressure Ryoke belt. The main deformation in the Sanbagawa belt is characterized by ductile flow in an E-W direction sub-parallel to the length of the belt, probably due to the oblique subduction. Kinematic studies suggest that the main ductile deformation caused a 35 - 65 % thinning of the region, which played an important role in causing exhumation of the region [30]. We cannot recognize any sedimentary sequence corresponding to forearc basin deposits near the Sanbagawa metamorphic belt. It may have thoroughly removed during exhumation of the metamorphic region.

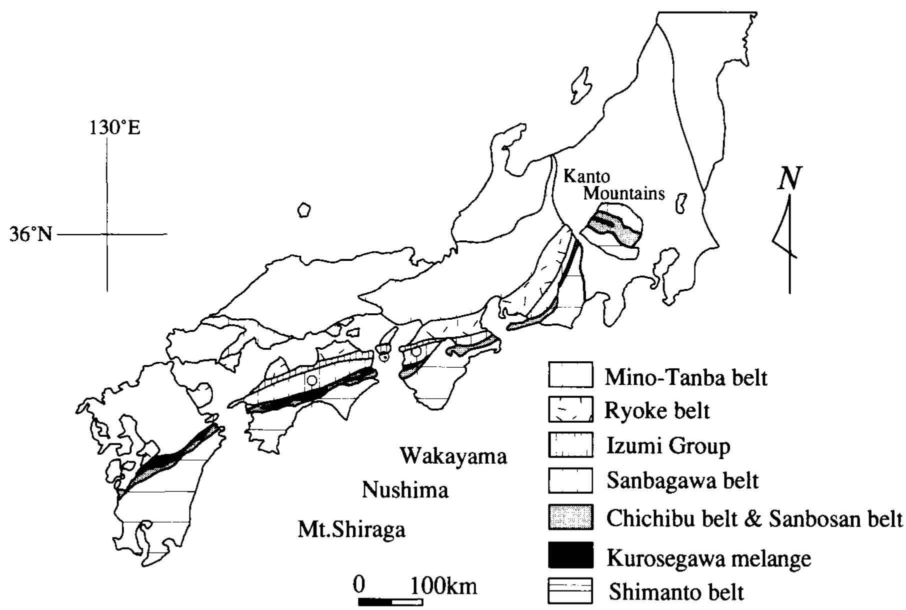

Figure 4. Geologic outline of southwest Japan (based on the data from Yamada et al. [42]).

In the Sanbagawa belt, ultramafic rocks are not common except for the Kanto Mountains of central Japan, and serpentinites are found only in the pelitic schists of the high-grade metamorphic zones. The occurrence of serpentinites in the Sanbagawa rocks at the Kanto Mountains has no relation to the metamorphic grade [31, 32]. Yano and Tagiri [32] explained the complicated thermal structure of this area in terms of the "shuffled-cards structure model" characterized by the repeated appearance of zone III (biotite zone) in the several sub-units. Irregular distribution of serpentinites may have resulted from the intense disturbance at the exhumation stage. The Sanbagawa belt lacks any serpentinite melange including exotic blocks. It seems likely that the Sanbagawa metamorphic rocks were formed under different environments from the Franciscan and Kamuikotan rocks.

4. POSSIBLE MECHANISMS FOR INCORPORATION OF SERPENTINITE INTO HIGH-PRESSURE METAMORPHIC TERRANES

4.1. Serpentinite diapir and proceeding submarine sliding

The blueschist-facies clasts obtained from Conical Seamount during ODP Leg 125 are small, less than 5 cm in diameter. On the other hand, tectonic blocks in serpentinite melanges of Franciscan and Kamuikotan terranes often range over a few kilometers in diameter. According to drilling records from the Conical Seamount, both drilling rate and degree of serpentinization of peridotites lessen gradually with the drilling depth; that is, higher density and larger blocks tend to be accumulated at the lower portion (Figure 5). Therefore, even if rising serpentinite grasped high-density and high-grade rocks at the deep portion, it could not bring them to the seafloor except for small fragments as seen at Conical Seamount. Serpentinite melanges in the Franciscan and Kamuikotan terranes are considered to represent a deep section of serpentinite diapir column similar to the Mariana serpentinites.

The serpentinites supplied by submarine gravity slides from the forearc seamounts into the trench could be again squeezed down to the deep subduction zone. Abundant serpentinites and serpentinized peridotites are exposed in the Franciscan and Kamuikotan terranes. They were commonly metamorphosed together with the surrounding rocks. In the Kamuikotan terrane, some serpentinites are found as lenses or layers in concordant with surrounding metasediments, and do not cut their bedding surfaces, suggesting sedimentary origin for these serpentinites [33]. At least some serpentinites in the Franciscan and Kamuikotan terranes appear to have been experienced a process from diapiric rise, through deposition at the trench by submarine gravity slide, to incorporation into the accretionary prism.

Serpentinite diapirism is one of the potentially important mechanisms for uplifting blueschist-facies blocks from subduction zones to forearc surfaces. This mechanism may have caused the recycling of blueschists as observed in the Franciscan and Kamuikotan terranes. It can also explain the occurrence of clastics of serpentinite and blueschist in the Great Valley Sequence and in the Yezo Supergroup, which are recognized as forearc basin sediments during the Franciscan and the Kamuikotan metamorphic events, respectively.

4.2. Scraping off hanging-wall peridotites

Serpentinites are less abundant in the Sanbagawa belt than those in the Franciscan and Kamuikotan terranes. As the belt totally lacks serpentinite melange including exotic blocks, it is possible that the Sanbagawa metamorphic rocks were not accompanied by serpentinite diapirism during metamorphism. As stated above, in many cases, serpentinites are found only in the pelitic schists of the biotite or garnet zone in the Sanbagawa belt ([34] and this study). For example, serpentinites occur only within the biotite or garnet zone at the eastern part of Wakayama City, Nushima at the southern end of Hyogo Prefecture, and at the western flank of Mt. Shiraga, central Shikoku (Figure 4). Small lenses and layers of tremolite-talc-chlorite rocks derived from serpentinites are often found in concordant with the foliation of surrounding pelitic schists near the serpentinite masses of these areas. Modes of occurrence of the tremolite-talc-chlorite rocks indicate that primary serpentinites were tectonically mixed with pelitic materials before the Sanbagawa metamorphism (Maekawa et al., in prep.).

A huge zone of diapiric serpentinite seamounts along the trench axis in the Izu-Ogasawara-Mariana forearcs suggests that the hanging-wall peridotites just above the subducting slab are pervasively serpentinized by water supplied from the pelagic sediments on the subducting plate. Whether the diapiric rise generate or not is probably depends on the state of stress field in the

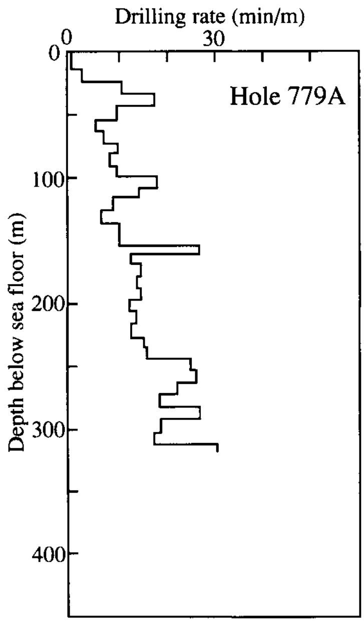

Figure 5. Relations between drilling rate and depth at the south eastern flank of Conical Seamount (Hole 779A, ODP Leg 125).

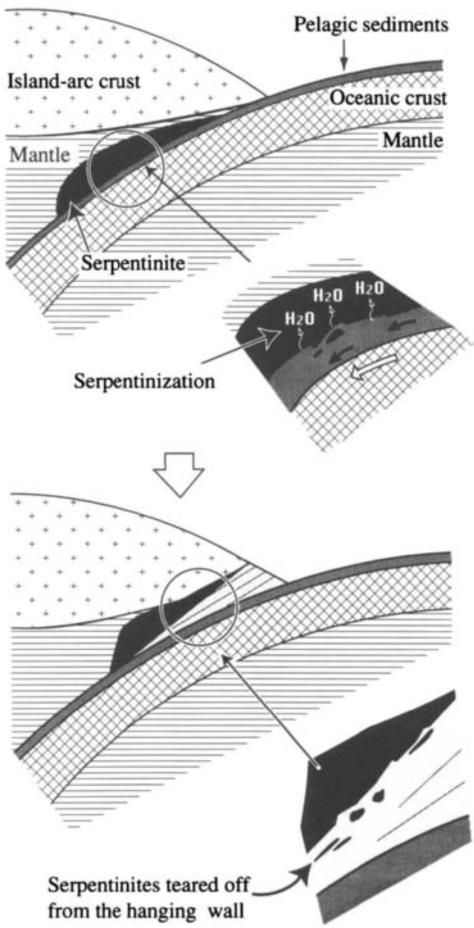

Figure 6. Tectonic model showing possible mixing of serpentinite with pelitic sediments within the accretionary prism.

forearc; the formation of up-rising serpentinite melange clearly favors a tensional stress field. Kunugiza et al. [34] clarified that some peridotites in the Sanbagawa belt were serpentinized before the Sanbagawa metamorphism. At least at the initial stage of subduction, subducting pelagic sediments encounter the hanging-wall peridotites, so it is highly probable that some of the serpentinite in the Sanbagawa metamorphic belt had been derived directly from the hanging-wall serpentinized peridotites (Figure 6). After the accretion of subducting materials began, subducting sediments cannot be in contact with hanging-wall peridotites because they were widely covered by the previously accreted materials. The initially accreted materials including serpentinites are expected to have been squeezed down to the deeper portion along the edge of high-temperature hanging-wall peridotites, and metamorphosed under higher-grade conditions in comparison with later accreted materials. The limited occurrence of serpentinites to the high-grade pelitic schists in the Sanbagawa belt support this hypothesis.

Serpentinized peridotites and serpentinites are commonly associated with mafic rocks in the Mikabu belt, which is situated at the southern end of the Sanbagawa belt. The genesis of these rocks were probably related to the collision of seamounts or oceanic plateaus above the subducting plate to the overriding plate, and are beyond the scope of this paper.

5. CONCLUDING REMARKS

Uyeda and Kanamori [35] classified the subduction zones into Mariana and Chilean types on the basis of the nature of the back arc regions. The former has back arc basins, whereas the latter has no back arc basins. They demonstrated that the difference depends on stress regime of arcs, that is, based on source mechanisms of intraplate earthquakes within the land-ward plates, present-day back arc areas are tensional for the Mariana-type arcs and compressional for the Chilean-type arcs. Nakamura [36] paid attention to the trench depth at the Honshu-Philippine-Pacific and the North America-Cocos-Caribbean triple junctions, and inferred that the abrupt change in the trench depth at the triple junction is due to the differential movement between two overriding plates. He demonstrated that an accretion of serpentinites in the forearc regions with tensional stress regime is one of the possibilities for explaining the existence of unstable triple junctions. The existence of serpentinites in the forearc regions is considered to reflect the tensional stress regime of the forearc, and serpentinite protrusion from depth to the seafloor is possible only in the forearc regions with tensional stress regime. On the basis of analysis of strain and stress distributions in the Sanbagawa belt, Toriumi [37] surmised that the high stress regime in the Sanbagawa belt require a subduction complex of Chilean-type.

Until now, a number of geologists have noted the difference in amount of serpentinites between the Sanbagawa-type and the Kamuikotan-type metamorphic terranes (e.g., Hashimoto, [38]). The Franciscan and Kamuikotan terranes are rich in serpentinites, and comprise serpentinite melanges including exotic blocks metamorphosed at the deeper portions in subduction zones. On the other hand, the Sanbagawa belt is rare in serpentinites and does not comprise serpentinite melange. The serpentinite melanges in the blueschist-facies metamorphic terranes are considered to represent the existence of serpentinite protrusion within the accretionary prism during metamorphism. The blueschist terranes with serpentinite melange were probably formed beneath the arcs with a tensional stress regime during the metamorphism, whereas those without serpentinite melange were formed beneath the forearcs with a compressional stress regime. The rare occurrence of serpentinites and lack of serpentinite melange in the Sanbagawa belt may have been caused by the compressional stress environment in the trench-arc region during the metamorphism. Under such compressional stress environments, scraping off hanging-wall peridotites seems to be only a possibility to incorporate mantle materials into the accretionary prism developing along the trench. The model can explain the limited occurrence of serpentinites to the high-grade pelitic schists in the Sanbagawa belt.

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Cretaceous environments of the Philippines

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Early geologic studies in the Philippines have habitually consigned all old looking rocks to the Cretaceous-Paleogene (KPg). However, the definite presence of Cretaceous units in the Philippines has only been confirmed recently by fossil and isotopic studies. Philippine Cretaceous rocks are exposed in the form of metamorphic basement rocks, crust-mantle sequences, magmatic/volcanic arcs and their associated sedimentary carapaces. Only selected areas with confirmed Cretaceous-dated formations are included in this paper, which occur in Palawan, southern Sierra Madre, Catanduanes, Cebu, Bohol, and Davao Oriental. For comparative purposes, some previously reported Cretaceous to Early Tertiary sequencesare included in this study. The Cretaceous complexes are generally capped by deep-water pelagic radiolarian cherts and limestones, which, in turn, are overlain unconformably by Early Tertiary rocks mostly Eocene in age. The change from Early to Late Cretaceous is from a relatively open and deep marine environment to shallow and landmass confined areas as evidenced by the presence of radiolarian cherts, pelagic foraminifers, and nannofossils to the appearance of *Orbitolina* spp., ammonites, rudists, palynomorph, and the occurrence of volcaniclastic rocks. The fossil assemblages further suggest a possible tropical environment all throughout the Cretaceous in the region. Available field data in terms of structure and geo-

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chemistry suggest that tectonic controls were dominant in effecting changes in the geologic setting during the Cretaceous in several areas in the Philippines.

1. INTRODUCTION

A study of known Cretaceous sequences in the Philippines was made with the purpose of determining the possible environmental changes that may have taken place in the Philippines during the Cretaceous [1]. Geologic studies in the past have designated all old looking rocks as Cretaceous-Paleogene (KPg). However, recent work involving biostratigraphic and isotopic studies have shown the presence of Cretaceous units in the Philippines. For this report, only those confirmed Cretaceous localities will be discussed and analyzed, and will focus on Palawan, southern Sierra Madre, Catanduanes, Cebu, Bohol, and Davao Oriental (Figure 1). Also included in this paper are selected Cretaceous to Early Tertiary sequences in the Philippines, whose characterization may be used for deciphering environmental changes that may have occurred from the Cretaceous to the Paleogene.

2. REGIONAL GEOLOGY

A complex system of subduction zones, collision zones and marginal seas characterize the current Western Pacific region, where the Philippines is located. The Philippine island arc system exposes tectonic features related to post-subduction, completed collision events and exhumed crust-mantle sequences. This island arc, which straddles the boundary of the Eurasian, Indo-Australian and the Philippine Sea Plates, is generally an assemblage of allochthonous arcs, ophiolite complexes, continental rocks and oceanic material welded together. These different terranes, the origin of which varies, were formed through subduction, collision or large-scale strike-slip faulting (Figure 1).

Structurally, the Philippine island arc system is bound on both sides by trenches [2, 3]. The Manila-Negros-Cotabato trench system defines the western boundary, while the East Luzon Trough-Philippine Trench is found on the east. Running along the whole length of the archipelago is the sinistral Philippine Fault, which accommodates whatever stress that the two binding trench systems cannot accommodate. Marginal basins, which include the South China Sea, Sulu and Celebes Seas to the west and the West Philippine Basin to the east also surround the Philippine archipelago.

The oldest known rocks in the Philippines are represented by the metamorphic formations

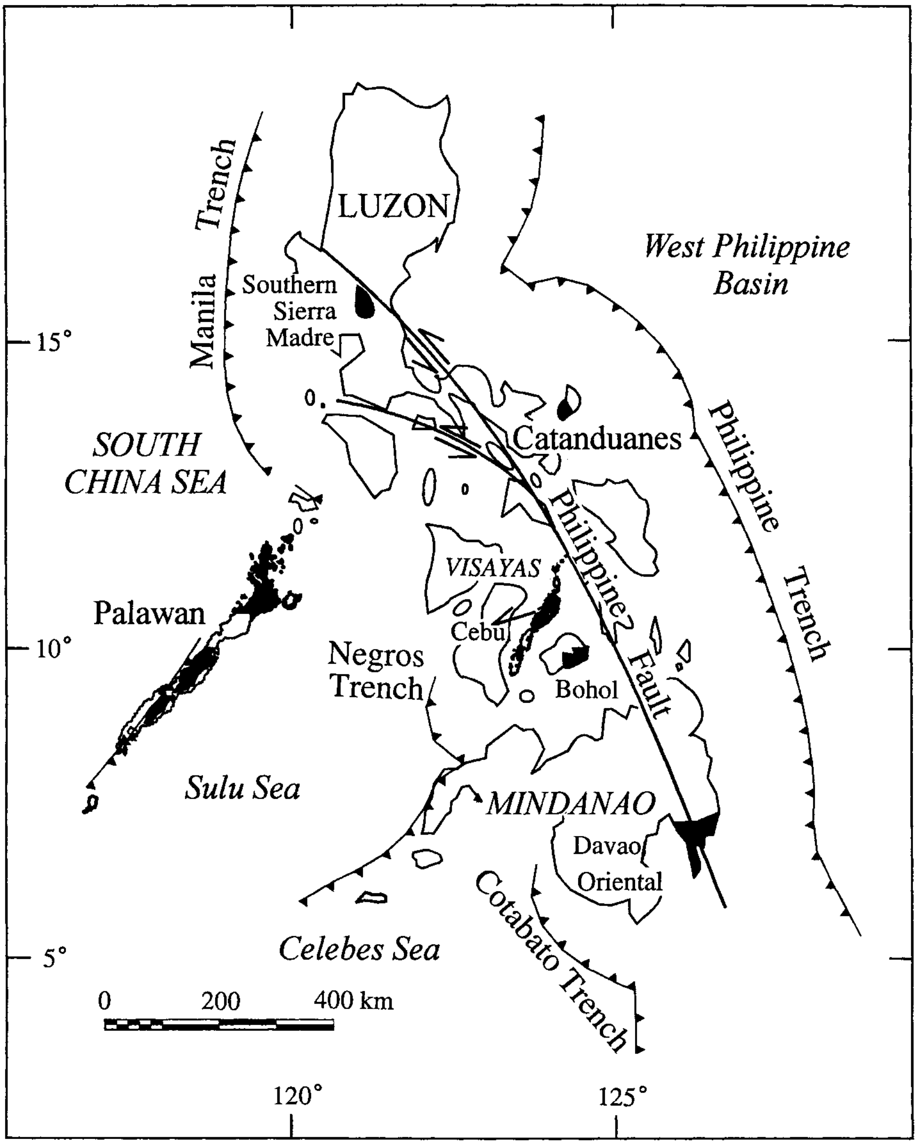

Figure 1. Map showing the distribution of representative exposures of Cretaceous-Tertiary sequences in the Philippines.

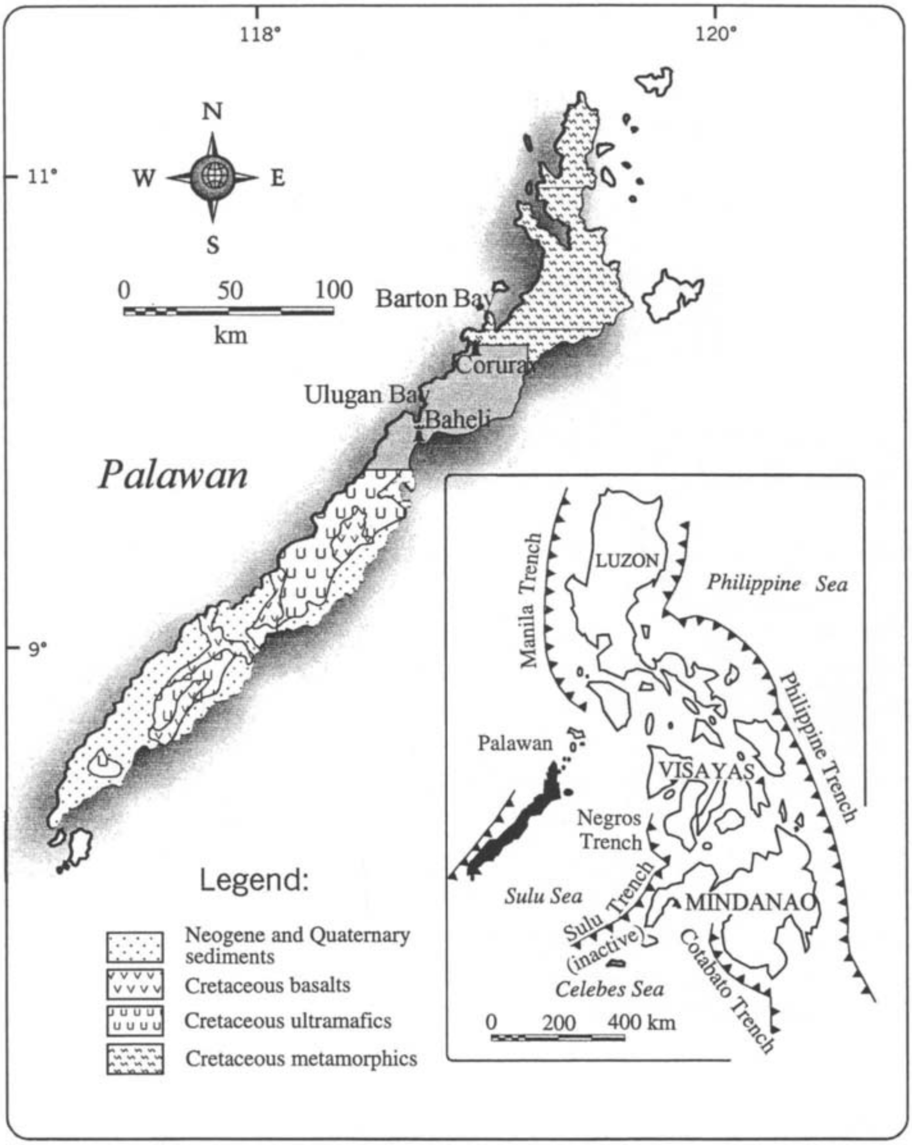

Figure 2. Simplified geologic map of Palawan. Inset is a location map of Palawan (modified after Mitchell et al.[8]).

located in north Palawan, Mindoro, Panay, and neighboring islands. These formations are related to the continental North Palawan block, that since late Miocene, collided with the Philippine Island arc system in the Mindoro-Panay area [4]. Datings of the pelagic sediment carapaces overlying ophiolite and ophiolitic complexes confirma pre-Tertiary genesis for some of these crust-mantle sequences [5, 6]. These ophiolite and ophiolitic complexes were thrust above younger Tertiary sedimentary sequences [4, 7, 8]. An understanding of the generation, evolution and emplacement of these ophiolite complexes is crucial in deciphering the geological evolution of the Philippine island arc system. Pre-Tertiary magmatic activities in the Philippines are represented by the dioritic rocks of Early Cretaceous age in Cebu Island [9] and similar rocks in Bohol Island [10, 11]. A number of other petrographically similar rocks are found in the archipelago but their ages are difficult to interpret due to unavailability of age data and the complexities of their tectonic settings.

3. CRETACEOUS - EARLY TERTIARY SEQUENCES IN THE PHILIPPINES

3.1. Palawan Island

Palawan Island is a northeast-trending ridge to the west of the Philippine Mobile Belt (PMB) (Figure 2). Balce et al. [12] and Holloway [13] considered North Palawan, Reed Bank, and Mindoro as a separate entity from the PMB based on stratigraphy and tectonic evidences. Likewise, Tamesis et al. [14] concluded that North and South Palawan, separated by the Ulugan Bay Fault, are two distinct terranes based on structural fabric and stratigraphy. UNDP [15] and BED [16] report contrasts between the Paleozoic and Mesozoic metasediments, granite intrusives and Tertiary sediments of North Palawan with the Mesozoic ophiolites and Tertiary melange noted in South Palawan. Pineda et al. [17] note, however, that the Sabang Thrust rather than the Ulugan Bay Fault is the trueboundary between North and South Palawan.

North Palawan belongs to the stable belt or the Palawan Physiographic Province [18] and includes the Busuanga Islands and the Cuyo Island. The Cretaceous and Early Tertiary sequence is represented by the Boayan Clastics, which is composed of alternating interbedded gray, micaceous, feldspathic sandstone and black, tuffaceous shale, radiolarian cherts and spilitic lavas [16, 19]. The formation is unconformably overlain by the Pabellion Limestone [19] (Figure 3). Based on the coccolith *Prediscosphaera cretacea,* the formation is dated Late Cretaceous [19].

The Palawan Ophiolite [20] comprises the Cretaceous sequence in central to southern Palawan (Figure 3). Peridotites, gabbros, metagabbros and local pillow lavas, radiolarites and red

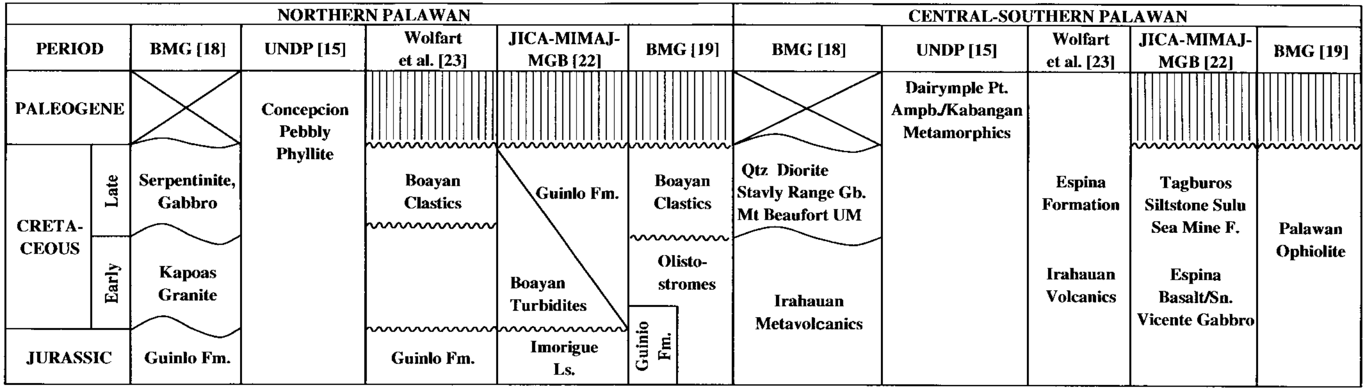

Figure 3. Comparative stratigraphy of Palawan (modified after BMG [19]).

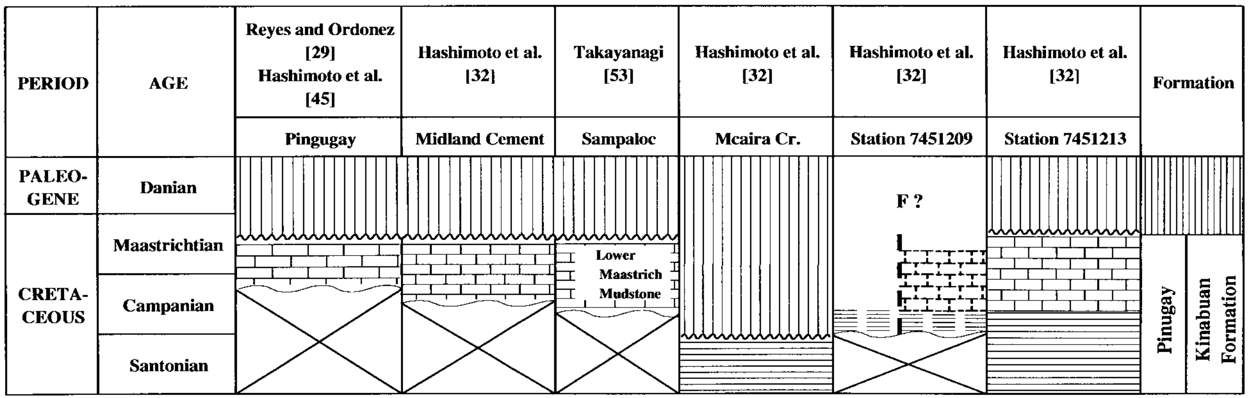

Figure 4. Comparative stratigraphy showing the faunal break between the Kinabuan and Maybangain Formations (modified from Hashimoto et al.[32]).

mudstones dismembered into several thrust slices make up this ophiolite body [17]. The ultramafic rocks, also termed the Mt. Beaufort Ultramafics [21] or the equivalent Ulugan Bay Ultramafic Complex [15] are composed of unaltered and serpentinized harzburgite, dunite and pyroxenite [19]. These are present in Mounts Beaufort and Malinao and the Ulugan Bay area. The ophiolite has a minimum age of Late Cretaceous based on radiolarian dating of cherts taken from inter-pillow matrix of the ophiolite [22].

The Bacungan River Group made up of the Maranat pillow lavas, Tagburos Siltstone and Sulu Sea Mine Formation [15] forms part of the Espina Formation. The JICA-MMAJ-MGB study [22] reported that the Maranat pillow lavas also contain basaltic tuff breccia. The Tagburos Siltstone is composed of massive, green siltstone and fine-grained sandstone. The Sulu Sea Mine Formation comprises alternating layers of red chert, ferruginous siliceous rock, conglomerate, sandstone and mudstone. A Late Cretaceous age is assigned to the Espina Formation on the basis of the *Tetralithus trifidus* assemblage [23], which suggests a late Campanian to early Maastrichtian age.

3.2. Southern Sierra Madre, Luzon

Southern Sierra Madre is host to two distinct rock groups of Cretaceous age. These are the Angat Ophiolite [24, 25] and the Kinabuan Formation (Figure 4). Scattered over the eastern portions of Rizal and Bulacan, and Nueva Ecija is the north-south trending Angat Ophiolite. On the other hand, the Kinabuan Formation is present in the Tanay and Antipolo areas of Rizal Province.

Arcilla [26] described the Angat Ophiolite as composed of layered and massive gabbros, sheeted diabase dikes, pillow basalts and sedimentary cover. In addition, Yumul and Datuin [25] reported that within the edges of pillow basalts are cherty material in-filling as well as basalt flows intercalated with red and greenish mudstones and cherts. High level gabbros, which are described as comprising abundant pegmatites with large hornblende crystals, crop out at the Puray River. A Turonian-Coniacian (early Late Cretaceous) age is assigned to the pillow lava-radiolarite unit directly below the Maybangain Formation along the Tayabasan River [26] (Figure 4). However, Haeck [27] came up with Paleocene and Eocene ages for radiolarians within the pillowed and massive basalts, which were considered by Encarnacion et al. [28] as the volcanic section of the ophiolite.

The Kinabuan Formation forms the sedimentary cover of the Angat Ophiolite and is composed of thin-bedded clastic rocks with interbeds of pelagic limestones in the lower member, while pelagic limestones make up most of the upper sequence. The lower member is further described by Haeck [27] as fine- to medium-grained calcarenite, buff to gray pelagic limestone and medium- to coarse-grained calcareous lithic to feldspathic arenite interbedded with black organic to light gray calcareous shale. The upper member is characterized by white to buff, light to dark pelagic limestones and minor light to dark gray calcarenite with rare interbeds of calcareous shale [19].

Reyes and Ordoñez [29] reported the occurrence of foraminifers ranging in age from Santonian up to early Maastrichtian in the Pinugay area (Figure 4). Likewise, Hashimoto et al. [31] reported Upper Cretaceous planktonic foraminifers *Globotruncana stuarti*, *G. lapparenti lapparenti*, *G. lapparenti bulloides*, and *G. gansseri* within the limestones below the intraformational conglomerate [29] in the Pinugay Hill. Reyes and Ordoñez [29] and Hashimoto et al. [32] recognized a faunal break ranging from upper Maastrichtian to Danian in the area (Figure 4). Furthermore, Upper Cretaceous radiolarians were reported in Tanay, Rizal by Ordoñez [30].

3.3. Catanduanes

Catanduanes is an island province located in the southeast of Luzon (Figure 1). It belongs to the Eastern Luzon Metamorphic Belt, which is considered a narrow basement sliver of an older arc crust or continent, rifted from its original setting by back-arc spreading. It also forms part of the Southeast Luzon Basin [33]. The comparative stratigraphy of the island is shown in Figure 5.

The Yop Formation consists of submarine andesitic flows, spilitic basalts, pillow lavas, intercalated cherts, arkosic and tuffaceous sandstones, andesitic volcaniclastic rocks and limestone [19, 34, 35] (Figure 5). Based on recent work, the Catanduanes Formation [18, 34] was found to be intercalated with the Yop Formation and, therefore, considered part of it [35]. The so-called Catanduanes Formation is composed of schists, volcaniclastic rocks, well-bedded sandstones and argillite with local interbeds of conglomerate [18, 34]. Based on the fossil

<table>
<tr><th>PERIOD</th><th>EPOCH</th><th>Crispin et al. [54]</th><th>Miranda and Vargas [34]
BMG [18]</th><th>Rangin et al. [55]</th></tr>
<tr><td>PALEOGENE</td><td></td><td></td><td></td><td></td></tr>
<tr><td rowspan="2">CRETACEOUS</td><td>Late</td><td rowspan="4">Hitona volcanic rocks
Cabugao Graywacke
Metavolcanic rocks
Metasedimentary rocks</td><td rowspan="2">Bonagbong Limestone
Yop Formation</td><td rowspan="4">Bonagbonag Limestone
Yop Formation</td></tr>
<tr><td>Early</td></tr>
<tr><td rowspan="2">JURASSIC</td><td>Late</td><td rowspan="2">Catanduanes Formation</td></tr>
<tr><td>Middle</td></tr>
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Figue 5. Comparative stratigraphy of Catanduanes Island (modified from BMG [19]).

content consisting of *Orbitolina* spp. and rare nannofossils such as *Micula staurophora* and *Eprolithus floralis*, the age given was Late Cretaceous (probably Turonian) [19, 35, 36].

The Codon Formation is a melange composed of volcanic and reworked limestone blocks in a volcaniclastic matrix (19). The Bonagbonag Limestone [37] associated with bedded calcareous siltstone, volcaniclastics and agglomerates was found to be enclosed in a fine-grained graywacke. The limestone is now considered as a megablock in the Codon Formation [19, 35]. Facies variation includes pebbly graywacke with limestone clasts to limestone breccias and megabreccias embedded in a graywacke matrix [19]. The graywacke matrix is barren of fossils, but the majority of the blocks contain Early Cretaceous *Orbitolina* spp. and Late Cretaceous *Globotruncana* spp., and unconformably overlain by the Eocene Payo Formation [19]. Based on stratigraphic relationships, the formation is given an age of Late Cretaceous to Early Paleogene [19].

Cretaceous intrusive bodies mainly composed of dolerite and gabbro are also encountered. Based on K-Ar dating, these range from 95.35 ± 5.7 Ma to 82.85 ± 2.6 Ma [38].

3.4. Cebu Island

The geology of Cebu Island consists of the northern-central and the southern parts of Cebu (Figure 6). A composite Cretaceous to Eocene stratigraphy, arranged in decreasing age, is as follows: the Tuburan Limestone, the Mananga Group (Pandan Formation and Cansi Volcanics), and the Lutopan Diorite.

The Tuburan Limestoneis a marbleized limestone sequence that contains Early Cretaceous fossils of *Orbitolina* spp. and Late Cretaceous *Rudista* spp. [40]. Porth et al. [41] refer to it as a pelletoidal micritic limestone containing fossils of pelecypods, algae, and foraminifers. The

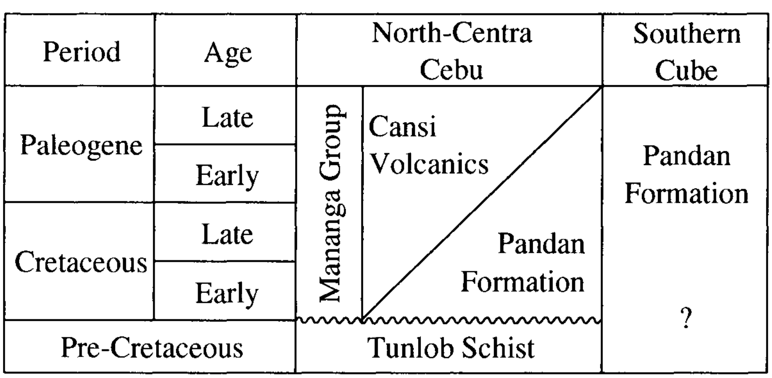

Figure 6. Stratigraphy of Cebu Island (modified after BMG [18] and JICA-MMAJ [39]).

Tuburan Limestone appears to be allochthonous occurring only as olistoliths within the Cansi Volcanics and as clasts in younger sedimentary rocks. This unit unconformably overlies the Jurassic (?) Tunlob Schist characterized by amphibolite schist and some phyllite and slates.

The Mananga Group refers to two distinct, coeval geologic formations, namely: the Pandan Formation and the Cansi Volcanics. Based on the occurrence of the Lower Cretaceous Tuburan limestone boulders within the Cansi Volcanics, the Mananga Group has been assigned to the Late Cretaceous to Early Paleogene age.

The type locality of the Pandan Formation is in Barrio Pandan, Naga, which is made up of metamorphosed limestone, shale, and conglomerate with lenses of lava flows [42]. The term Pandan Series is also used for the sequence of sedimentary and volcanic rocks in the southern portion of Central Cebu and in the Northern Highlands [42]. They referred to this series as consisting of a sequence of sedimentary rocks with locally alternating layers of lava flows. The sedimentary sequence consists of thermally metamorphosed thin- to medium-bedded sandstone, thin-bedded shale and chert, and occasional beds of conglomerate and limestone. Porth et al. [41] described the Pandan Formation as composed only of greenish-gray, calcareous, partly siliceous siltstone and light gray, silty, *Globotruncana*-bearing limestones. *Globotruncana* spp. puts a Late Cretaceous age for the limestone and thus suggests a Late Cretaceous to Early Paleogene age for the Pandan Formation [18].

The Cansi Volcanics are essentially composed of massive lava flows, breccias, and polymictic agglomerates [41, 43]. The lava flows are characterized by slightly metamorphosed basaltic to andesitic rocks. Textures vary from aphanitic, fine- to coarse-grained porphyritic to amygdaloidal. Common amygdules include quartz, calcite, and chlorite. The agglomerates contain pebble- to boulder-size fragments of igneous and sedimentary rocks. Pillow structures were found in basalt and andesitic lavas. The *Orbitolina*-bearing limestone boulders and the *Rudista*-bearing limestone lenses [40] within the Cansi Volcanics were dated Late Cretaceous. This places the Cansi Volcanics as younger than Late Cretaceous.

The Lutopan Diorite generally refers to diorite stocks commonly found in the highlands of northern-central Cebu [18]. The Lutopan Diorite varies in composition from hornblende diorite to hornblende quartz diorite to quartz diorite [18] and is generally characterized by having a medium- to coarse-grained, granular texture, made up of oligoclase (or andesine), quartz, hornblende and biotite, and occasionally chlorite. Accessory minerals include magnetite, apatite, and zircon [39]. Common alterations observed in the Lutopan Diorite consist of silicification, sericitization, pyritization, argillization, and epidotization found especially in the vicinity of the Atlas Mine District. Several K-Ar dating on the Lutopan Diorite yields a Paleocene age [3, 9, 39]. However, Walther et al.[44] determined an Early Cretaceous age for the Lutopan Diorite.

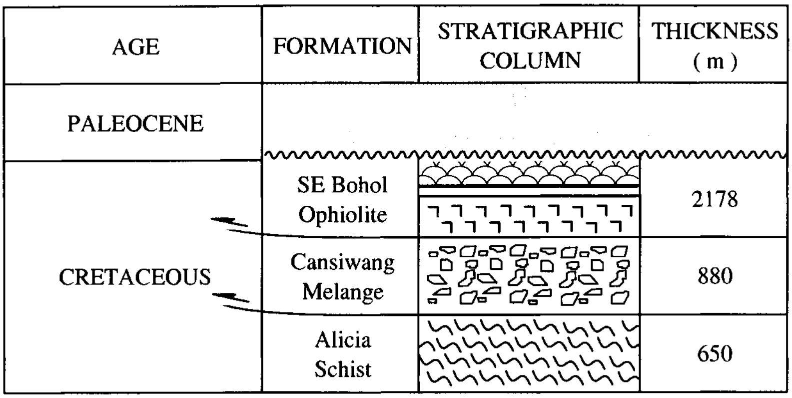

Figure 7. Stratigraphy of Southeastern Bohol (modified after NIBS Geology 170 Class [56]).

3.5. Bohol Island

The island of Bohol is situated in the Visayan Basin, and its structure and stratigraphy are comparable with those of Cebu, Negros, Panay and Leyte [18, 41]. The stratigraphy of Southeastern Bohol is shown in Figure 7.

The Alicia Schist comprises the oldest stratigraphic unit in Bohol Island and is a metamorphic body composed of sericite schist, quartz-sericite schist and amphibolite schists [45, 47]. This metamorphic facies occurs as NW-SE trending belts in the northeast portion of the island. The Alicia Schist, based on recent works, was identified as having been formed through regional metamorphism and is no trelated to the Southeast Bohol Ophiolite Complex emplacement [48, 49]. This new interpretation is based on the recognition of a melange body that is thrust on top of the Alicia Schist and the lack of a regressive metamorphic gradient away from the ophiolite unit [48]. From field observations, the protoliths of the metamorphic rocks are sandstones and basalts. Although no fossils are reported from the Alicia Schist, structural as well as stratigraphic relationships with other lithologies suggest a Cretaceous age for the Alicia Schist [46, 50].

The Cansiwang Melange, a newly recognized melange unit, crops out as NE-SW trending bodies along the eastern portion of the island. It is an ophiolitic melange characterized by sheared fragments of basalts, harzburgites and minor clastic rocks and cherts in a serpentinite matrix [48]. This melange body represents an accretionary prism deposit that may have been formed in an incipient subduction process [48]. The Cansiwang Melange is thrust above the Alicia Schist and below the Southeast Bohol Ophiolite Complex. The serpentinitic matrix is barren of fossils. However, cherts (Cretaceous?) occur as clasts in the melange. Based mainly on stratigraphic relationship, the Cansiwang Melange is given a Cretaceous age [48].

The Southeast Bohol Ophiolite Complex is composed of residual peridotites, layered clinopyroxenites, dunites and gabbros, massive gabbros, sheeted dike complex and pillow basalts capped by a sedimentary carapace and is exposed mainly as NE-SW trending belts in Southeast Bohol [46, 50]. The complete ophiolite sequence is exposed along the Alejawan River west of Duero, while segments of the ophiolite can be found west of the towns of Guindulman and Alicia. The ophiolite is thrust above the Cansiwang Melange and unconformably overlain by middle Miocene clastic and carbonate rocks. Intercalated with the pillow basalts of the ophiolite are either tuffaceous sediments or bedded cherts. Preliminary results of age dating on the cherts, as well as the stratigraphic position of the ophiolite suggest a Cretaceous age of formation.

Cretaceous ages of some intrusives in Bohol Island are based primarily onthe correlation done between the Central Cebu intrusives such as the Matugan and Talisay-Minglanilla Diorites and the Talibon Diorite. Porphyritic diorites from the Atlas Mine in Central Cebu are dated to be 108 ± 1 Ma [44].

3.6. Davao Oriental

The early Late Cretaceous-Eocene stratigraphy (Figure 8) of the Mati-Northern Pujada Peninsula, Davao Oriental, is characterized by the Pujada Ophiolite and its sedimentary cara-

PERIOD	EPOCH	FORMATION	MEMBER	STRATIGRAPHIC COLUME
CRETACEOUS	Late	Iba Formation	Dawan Clastic Rocks	
			Bitanagon Limestone	
			Lampasan Chert	
	Early Late	Pujada Ophiolite	Pillow Basalt	
			Gabbro	
			Harzburgite	

Figure 8. Stratigraphy of Mati, Northern Pujada Peninsula (modified from Queano et al.[52]).

pace, the Iba Formation. The Pujada Ophiolite can be found in the Pujada Peninsula, its type locality, while the components of the Iba Formation are located northeast of the Peninsula.

The Pujada Ophiolite is considered a complete ophiolite sequence consisting of serpentinized harzburgites, massive and layered gabbros, sheeted dike complexes and pillow basalts. The harzburgites, composed essentially of olivine and orthopyroxene and accessory minerals such as epidote and calcite, are characterized by moderate to intense serpentinization and are cut by a series of generally west-verging thrust faults. The massive and layered gabbros are generally medium- to coarse-grained in texture and are essentially made up of oligoclase and augite and minor chlorite. The layered gabbros follow a northwest-southeast trend, while the sheeted dike complex is found in the Oregon-Upper Anitap area in Monserrat. The pillow basalts cover an extensive area in Mati and in Northern Pujada Peninsula. Smaller outcrops also occur in Guanguan Peninsula. In Barangay Badas, the pillow basalts were observed to be locally intercalated with the chert layer. The altered equivalents of the pillow basalts were found in Camansi near Mati. An early Late Cretaceous age is assigned to this formation based on its sedimentary carapace discussed below.

The Iba Formation is the sedimentary carapace of the Pujada Ophiolite. It has three members: namely, the Lampasan Chert, the Bitanagon Limestone, and the Dawan Clastic Rocks. The Lampasan Chert is a massive to bedded formation, locally present as recrystallized lenses. It conformablyoverlies the pillow basalts as observed in Purok Lampasan. The presence of the foraminifers *Globotruncana calcarata, G. stuarti, G. calciformis-contusa* indicates a Late Cretaceous age. The Bitanagon Limestone found in the Bitanagon River is either massive or thinly bedded with average bed thickness of 10 cm. Bedded limestone observed in Bitanagon has an average bed thickness of 10 cm. The bedded limestone observed in the Bitaogan River is in fault contact with the phyllites. This fault contact has a northwest strike and dips to the northeast. Stylolitization can be observed in some outcrops. The presence of limestone lensoids in the Lampasan Chert indicates a conformable relationship between these units. The Dawan Clastic Rocks constitute the uppermost member of the Iba Formation. It consists of alternating beds of poorly-sorted, greenish gray sandstone and reddish-brown shale. The clasts are arkosic in nature. These exposures are observed along the Badas Road.

The degree of metamorphism decreases from the southern part of the Peninsula to its northern part and was not observed east of Dawan. Phyllites and green schist-facies metamorphic rocks have their type locality in the Bitaogan River. Protoliths range from the ultramafic-mafic components of the ophiolite basement complex to the Bitanagon Limestone of the Iba Formation. Age of metamorphism may be dated as Early Tertiary, since this is only observed in the Pujada Ophiolite and members of the Iba Formation.

4. SUMMARY AND CONCLUSIONS

Tectonic reconstructions in North Palawan suggest that sedimentation in the area occurred mostly in the southeastern margin of mainland Asia during Late Cretaceous to Paleogene [16]. The deposition took place in a deep marine environment based on radiolarian cherts and spilitic lava deposits. The presence of quartzose wacke and feldspathic sandstone indicates a continental source.

In southern and central Palawan, the Cretaceous unit is the Palawan Ophiolite, which was probably derived either from the proto-Celebes Sea Basin or the proto-South China Sea Oceanic Crust. The ophiolite together with the radiolarites and red mudstones indicates a deep marine environment possibly below the carbonate compensation depth (CCD).

The Angat Ophiolite of southern Sierra Madre is an allochtonous terrane, that was accreted to the Northern Luzon block, and formed in a back-arc environment [24]. Geochemical and sedimentary studies by Yumul and Datuin [25] suggest subduction-related marginal basin environment of formation. The ophiolite may have been emplaced before or during the Cretaceous as denoted by the presence of Cretaceous foraminifers on its sedimentary cover. A deep marine setting is noted by the pelagic deposits, which characterize both the lower and upper members of the Kinabuan Formation. An erosional event occurred between the Late Cretaceous and the Paleogene, as manifested by the presence of a faunal break between the Kinabuan and Eocene Maybangain Formations.

Sedimentation during the Cretaceous was apparently localized in Catanduanes. Spilitic lavas, cherts, clastic rocks and carbonates suggest a marine, probably neritic to bathyal deposition marked by a period of intense volcanic activity [33]. Based on the faunal content and petrographic characteristics of the carbonate rock samples taken in the area, Fernandez et al. [36] made some generalizations. These include the following: (a) during the late Early Cretaceous time, the area was probably shallow, as indicated by the presence of *Orbitolina* spp.; and (b) the abundance of planktonic foraminifers and nannofossils in the Late Cretaceous samples suggests deepening of the area due to either marine transgression or subsidence.

The Early Cretaceous of Cebu Island saw shallow marine deposition of the *Orbitolina*-bearing Tuburan Limestone above the Tunlob Schist. The transition between the Tunlob Schist and Tuburan Limestone is marked by an unconformity. The sequence of sedimentary (chert, sandstone, shale, and conglomerate) and volcanic rocks represented by the Mananga Group suggests a marine, probably neritic deposition during the Late Cretaceous to Paleogene based on rudists. This, however, was preceded by an erosional event, as indicated by the presence of the Tuburan Limestone boulders within the Cansi Volcanics. Based on the intercalating relationship between the sedimentary sequence and the lava flows in the Pandan Formation, volcanic events probably occurred at irregular intervals during the Cretaceous to Paleogene. Volcanic activity was also correlated with emplacement of the Lutopan Diorite and the coeval formation of the Atlas porphyry copper deposit during the Early Tertiary.

The Cretaceous-Tertiary evolution of Bohol Island is marked by major tectonic activities. The presence of the Cansiwang Melange, which represents an accretionary wedge, suggests incipient subduction. This subduction may have occurred along the paleo-Southeast Bohol Trench [8, 46] and led to the subsequent docking or underthrusting of the regionally metamorphosed Alicia Schist [50]. Subduction continued along this paleo-subduction zone until the Southeast Bohol Ophiolite Complex was onramped from the northwest [46, 50]. The magmatism related to this pre-Tertiary subduction is reflected by the intrusives in Central Cebu.

The presence of limestone and chert exposures in Davao Oriental would suggest a deep, open-sea depositional environment. Petrographic and geochemical analyses of the basalts indicate a mid-oceanic ridge or back-arc basin affinity. Furthermore, the presence of chert rinds and intercalation in the basalts indicates a syngenetic relationship. Thus, it can be inferred that the pillow basalt, and the rest of the ophiolite sequence formed either in a back-arc or mid-oceanic ridge setting. This problem could be resolved by looking at the Dawan Clastic Rocks. Volcanism was then active during the Late Cretaceous to Paleogene, as indicated by the arkosic nature of the clasts in this unit. The clasts are poorly sorted suggesting a proximal source. These evidences would point to a back-arc basin origin. Hawkins et al. [51] favor this tectonic setting largely due to the large arc material component.

The Philippine Cretaceous complexes are generally capped by deep-water pelagic radiolarian cherts and limestones, which, in turn, are unconformably overlain by Early Tertiary rocks mostly Eocene in age. Apparently, the change from Early to Late Cretaceous was from a relatively open and deep marine environment to a shallow and landmass confined regions, as indicated by the presence of radiolarian cherts, pelagic foraminifers and nannofossils to the appearance of *Orbitolina* spp., ammonites, rudists, palynomorphs and the occurrence of volcaniclastic rocks. The fossil assemblages further suggest a possible tropical environment throughout the Cretaceous.

In the interpretation of the Cretaceous sequences studied, it appears that the region had undergone different evolutionary histories during the period. The Philippine island arc system is a composite terrane made up of varied source rocks and developmental processes that were accreted/amalgamated together through subduction-related collisions and large-scale strike-slip faulting.

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Cretaceous Environments of Asia
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Cretaceous environments in Viet Nam, Laos and Cambodia

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In the Indochina Peninsula the sedimentation during the Cretaceous took place under the continental conditions. In some basins were formed normal continental red beds, but in some others volcanogenic formations. During the Cretaceous the physical change of the studied areas gave rise to planation of the relief to form large lakes or dead seas with brackish or salt water. The climate became hotter and drier in the Late Cretaceous that influenced strongly the existence of organisms. In some tectono-volcanic depressions, igneous extrusions were active especially in the Late Cretaceous.

1. OUTLINE OF THE CRETACEOUS SYSTEM

1.1. Distribution

In the Indochina Peninsula the Cretaceous System is represented by two sequence types: 1) red continental beds, in some places bearing evaporites; and 2) volcanogenic formations with interbeds of red continental sediments at the basal part.

Cretaceous continental formations are widely distributed in Middle and South Laos and Southwest Cambodia. They are met also in narrow bands in North Viet Nam, and in some islands of the Gulf of Thailand. At the same time, Cretaceous volcanogenic formations are distributed only in tectono-magmatic activated depressions of Viet Nam, which are located in the north and in the south (Figure 1).

1.2. Geologic setting

In Laos, continental red beds have been found in two basins: the Vientiane Basin, which is a part of the Sakon-Nakhon Depression of Thailand, and the Donghen Basin, which is a part of the Khorat Depression also of Thailand (Figure 1).

In the Vientiane Basin, Cretaceous sediments have been divided into the Lower Cretaceous Cham Pa Formation, consisting of conglomerate, sandstone, red-brown siltstone and white arkosic sandstone, 400 m thick; and the Upper Cretaceous Tha Ngon Formation, composed of red-brown and white claystone, sodium salt, gypsum and potassium salt diapirs, passing upward to red-brown siltstone and calcareous claystone, 434 m thick [1].

In the Donghen Basin, Cretaceous sediments have been described in the Donghen Group including four formations: 1) the Upper Jurassic Nam Phouan Formation, composed of basal

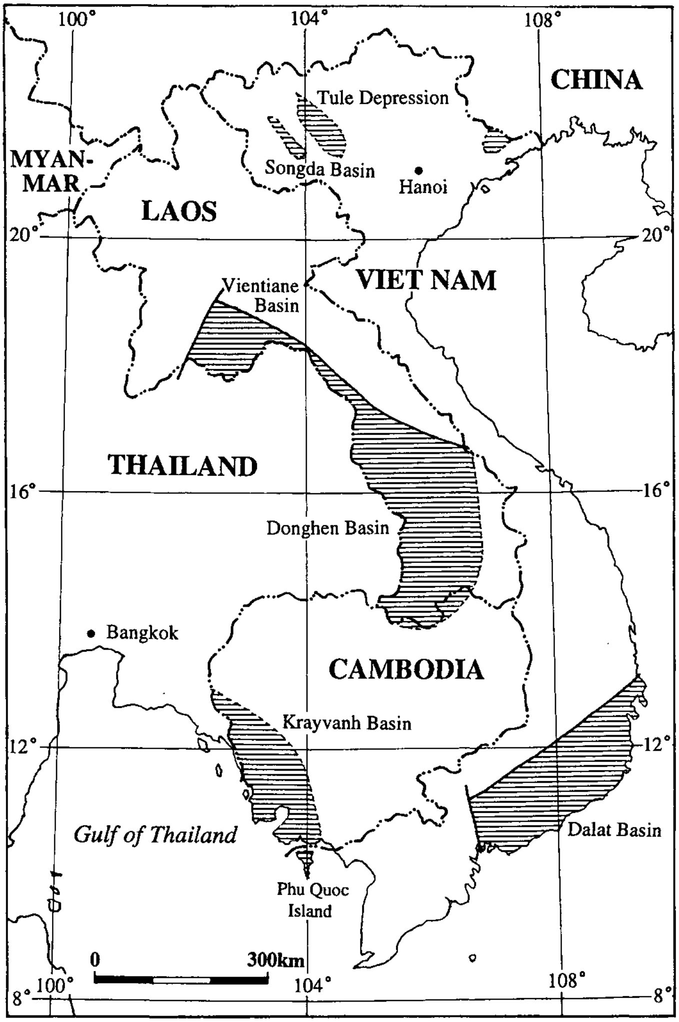

Figure 1. Distribution of Cretaceous in Viet Nam, Laos and Cambodia. Thick line: fault

conglomerate, grey-greenish copper-bearing (chalcopyrite, bornite) sandstone and interbeds of red-brown siltstone, 350-400 m thick; 2) the Lower Cretaceous Nam Xot Formation, composed of light-coloured, cross-bedded sandstone and gritstone with interbeds of red-brown siltstone bearing Early Cretaceous sporomorphs, locally bone fossils and footprints of different dinosaurs, and brackish-water bivalve shells, 500-550 m thick; 3) the Lower-Middle Cretaceous Nam Noy Formation, composed of red-brown siltstone and claystone with interbeds of micaceous sandstone, 450 m thick. On the bedding surface ot siltstone there are dessication cracks; 4) the Upper Cretaceous Noong Bua Formation, including fine-grained sandstone, red-brown siltstone with interbeds of gypsum, sodium salt (85-251 m thick) and anhydrite, 650-700 m thick [field data of Tran Ban, Vu Khuc and Vu Chau, 1998]. The Donghen Group is the facies equivalent to the Khorat Group of Thailand.

In Southwest Cambodia and some Vietnamese islands in the Gulf of Thailand (Phu Quoc and Tho Chu), Cretaceous continental red beds have been described as the "Upper Sandstone" or the Phu Quoc Formation with the following rhythmic sequence: 1) red-brown claystone grading upward to pinkish sandstone, 200 m thick; 2) red-brown claystone and sandstone, 160 m thick; 3) red-brown claystone grading upward to white arkosic sandstone with interbeds of conglomerate, 250 m thick; 4) red claystone, white arkosic sandstone, 300 m thick; and 5) pinkish, yellowish sandstone, 500 m thick. The total thickness of the formation reaches 1,410 m. Sediments of the lower part of this formation furnish sporomorphs and leaf imprints of Early Cretaceous age (see Table 1) [2, 3].

In Northwest Viet Nam, Cretaceous red beds have been described as the Yen Chau Formation, composed of basal conglomerate, gritstone, sandstone grading upward to red-brown silstone and claystone with interbeds of sandstone, 650-800 m thick. Sandstone of the lower part furnishes rare remains of fresh-water bivalves, meanwhile siltstone contains Cenophytic leaf imprints that date the formation as the Late Cretaceous. On the bedding surface of siltstone and sandstone there are ripple marks and dessication cracks [1].

Cretaceous volcanogenic formation in Northwest Viet Nam are distributed in the Tule Depression. They have been divided into two units: the Van Chan Formation and Ngoi Thia Formation.

The Van Chan Formation comprises basal conglomerate, gravelstone, siltstone, tuffaceous sandstone and siltstone, grading upward to rhyolitic tuffs, quartz orthofelsite with interbeds of coaly shale bearing leaf imprints, 2700-3000 m thick. The formation has been supposedly dated as Late Jurassic-Early Cretaceous. The Ngoi Thia Formation comprises porphyritic rhyolite, quartz porphyry, comendite with interbeds of their tuffs, 1700-2000 m thick. Isotopic dating has given a value of 79 Ma, corresponding to Late Cretaceous [3].

In South Viet Nam volcanogenic formations are widespread in the Mesozoic Dalat Depression, and include the Nha Trang and Don Dzuong Formations.

The Nha Trang Formation is composed of basal conglomerate, tuffaceous sandstone, interbeds of red-brown siltstone, andesite, andesito-dacite, dacite and their tuffs, grading upward to rhyolite, trachyrhyolite, porphyritic felsite and their tuffs, 500-550 m thick. Siltstone of the lower part yields palynomorphs of Cretaceous age (see Table 1).

The Don Dzuong Formation consists mainly of volcanics with porphyritic dacite, porphyritic felsite, rhyodacite, rhyolite, porphyritic rhyolite and their tuffs with basal conglomerate, tuffaceous sandstone and red-brown tuffogenous siltstone in the mid of the sequence, 1200-1300 m thick. Siltstone of the middle part furnishes Cretaceous palynomorphs (see Table 1) [3].

2. ENVIRONMENTS

Geological evidence shows that during the Cretaceous in the Indochina Peninsula there existed two types of depositional basins. In the first basin type, the depositional sites took place under stable continental conditions, forming normal continental beds of red colour. The

Table 1
Stratigraphic correlation of Cretaceous formations in Viet Nam, Laos and Cambodia

<table>
<tr><th rowspan="2">Series</th><th colspan="2" rowspan="2">NORTH VIET NAM</th><th rowspan="2">CENTRAL VIET NAM</th><th colspan="3">LAOS</th><th rowspan="2">CAMBODIA</th></tr>
<tr><th>Vientiane Basin</th><th colspan="2">Donghen Basin</th></tr>
<tr><td rowspan="2">UPPER CRETACEOUS</td><td rowspan="2">Yen Chau Fm.
Red beds bearing Unio sp., Diospyros, Ulmus, Gyraulus.</td><td rowspan="2">Ngoi Thia Fm.
Rhyolite, comendite, tuffs.</td><td rowspan="2">Don Dzuong Fm.
Rhyodacite, rhyolite, porphyritic felsite, tuffs with Lygodium, Picea, Cedrus.</td><td rowspan="2">Tha Ngon Fm.
Sodium and potassium salt-bearing sediments with Exesipollenites, Classopollis.</td><td rowspan="4">Donghen Group</td><td>Noong Bua Fm.
Sandstone, siltstone, gypsum, salt, anhydrite</td><td rowspan="4">Upper Sandstone Fm.
Red beds bearing Gleichenoides, Sphenopteris, Leckenbya, Concavisporites, Diapites.</td></tr>
<tr><td rowspan="2">Nam Noy Fm.
Siltstone, claystone, interbeds of sandstone</td></tr>
<tr><td rowspan="2">LOWER CRETACEOUS</td><td rowspan="2"></td><td rowspan="2">Van Chan Fm.
Quartz orthophyre, tuffs, red beds bearing Coniopteris, Nilssonia, Pterophyllum, Podozamites.</td><td rowspan="2">Nha Trang Fm.
Andesite, dacite, tuffs with Selaginella, Seitotylus, Taxodium, Lygodium.</td><td rowspan="2">Cham Pa Fm.
Red beds</td></tr>
<tr><td>Nam Xot Fm.
Sandstone, gritstone, interbeds of siltstone with dinosaurs, Plicatounio, Hoffetrigonia, Nippononaia.</td></tr>
</table>

second basin type reflects continental volcanism, that formed thick volcaniclastic deposits bearing thick members of effusives.

The first basin type is located in Middle and Lower Laos, Southwest Cambodia, North Viet Nam and in some islands of South Viet Nam. In these basins, one can observe some changes during the Cretaceous.

During the Early and Middle Jurassic, all these basins were submerged, but by the end of the Jurassic time, the sea regressed, establishing continental conditions in these areas. During the Late Jurassic and the Early Cretaceous the marine low-stand gave rise to mainly coarse-grained, cross-bedded fluvial deposits. The climate in these basins is presumed to have been warm and humid. The vegetational cover supported an assemblage of leaf-eating dinosaurs with different genera of sauropods, ornithopods, theropods and iguanodontids [4]. In rivers, there were turtles and *Unio*, meanwhile in lakes there were brackish-water bivalves, such as *Plicatounio, Hoffetrigonia* [5], *Nippononaia*, etc.

This environmental pattern changed in the Late Cretaceous. The topography got matured, wherein large lakes formed. Upper Cretaceous deposits are composed of claystone and siltstone of lacustrine facies, bearing thick members of evaporites, such as sodium salt (up to 85-251 m thick), gypsum anhydrite and potassium diapirs, suggesting a hot and dry climate [6, 7]. Rare fossils have been found in Upper Cretaceous sediments, mainly of palynomorphs.

The above-described features are fully developed only in the large basins of Laos. In the basins of Viet Nam and Cambodia, that may have been narrow intermontane rift basins, the sediments are mainly of alluvial and proluvial origins with rare and very thin interbeds of evaporites in the upper part of the sequence. The change in environment is difficult to be recognized, except the rarity of fossils in the upper part of the sequence. In Lower Cretaceous beds have been found leaf imprints, such as *Gleichenoides stenopinnula, G. pantiensis, G.* cf. *maranensis, Sphenopteris mantelli, Leckenbya valdensis* of Neocomian age, together with such palynomorphs as *Concavisporites* sp., *Diapites spinosa*, *Cicatricososporites* sp. of Barremian-Aptian age [2], and silicified woods such as *Prototaxoxylon asiaticum, Proto-podocarpoxylon orientale,* and*P. paraorientale* of Cretaceous age [8].

In tectono-volcanic basins located in Northwest and South Viet Nam deposition during the Cretaceous was also continental. Initially normal deluvial, proluvial continental red sediments were followed by volcaniclastic beds, reflecting volcanic activity throughout the Cretaceous. First deposited were intermediate rocks, then felsics, and finally felsic and alkaline rocks. In the late phases, tuffaceous rocks became rarer, ceding the place to volcanics. This reflects strongest activities of volcanoes at that time.

In these volcanogenic formations fossils are very rare. From Lower Cretaceous beds of some localities have been found such palynomorphs as *Selaginella* sp., *Taxodium* sp., *Seitotylus* sp., *Lygodium* sp., *Picea* sp., *Cedrus* sp., *Schizosporites* sp., etc. [1].

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Cretaceous environments of northeastern Thailand

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The nonmarine Cretaceous rocks are widespread in the northeastern part of Thailand. They belong to the Phra Wihan, Sao Khua, Phu Phan and Khok Kruat Formations of the Khorat Group and the overlying Maha Sarakham Formation. Reddish-brown to light-gray sandstones, conglomeratic sandstones, siltstones, claystones and clongomerates are the main lithologies of these rocks; calcrete nodules and silcretes are also present in claystones but salts and gypsum are found only in the Maha Sarakham Formation. The rocks are interpreted as having been deposited by the meandering and braided rivers in semi-arid to arid conditions. Age determinations are based mainly on vertebrates, bivalves and palynomorphs, indicating that the rocks are reassigned to the late Early to Late Cretaceous. Previously known lime-nodule conglomerates and silicified wood-like beds and nodules have been re-interpreted as paleosols and silcretes, respectively. These paleosols and silcretes are widespread in maroon to reddish-brown claystones of the Khorat Group particularly in the Sao Khua Formation. Petrographically and geochemically, paleosols and silcretes are similar to those in semi-arid to arid climate silcretes in South Africa, Canada and other regions. As a result, paleosols and silcretes may be used as paleoclimate indicators for the Cretaceous rocks, indicating that the rocks were deposited in the meandering fluvial system during semi-arid paleoclimate.

According to the physical, biological and climatological changes in the Cretaceous rocks of northeastern Thailand, there have been the repetitive changes from meandering to braided river systems corresponding to paleoclimatological changes from semi-arid conditions in the Early Cretaceous to the arid paleoclimate in the Late Cretaceous. Biological changes are depending on physical and climatological changes, in which the faunal abundance and diversity are confined to rocks deposited by the meandering river system in semi-arid paleoclimate.

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1. INTRODUCTION

The Cretaceous sedimentary rocks of Thailand consist mainly of nonmarine facies. Although brackish-water to nonmarine conditions occur in a few places in the southern peninsula, which will not be presented here, the marine rocks have not yet been reported in Thailand. The nonmarine Cretaceous is widespread mainly in the northeastern part of the country and is also found in the small area in the south. On the Khorat Plateau in northeastern Thailand, a sequence of Permian carbonates and shallow siliciclastics is unconformably overlain by predominantly fluvial/lacustrine Mesozoic clastics of the Triassic Huai Hin Lat Formation, the Late Cretaceous Khorat Group and the Cretaceous Maha Sarakham Formation. The Cretaceous of the Khorat Group was reported as early as the 1960's, but most paleontologists began to study them in the 1980's. Bivalves were the first studied by Japanese workers (e.g. Kobayashi et al. [1]). Other fossils reported are vertebrates, invertebrates and palynomorphs.

The scope of this paper is to synthesize the changes in physical, biological and climatological aspects of the Thai Cretaceous and the impacts of those changes on the total environments. Although the climatological environments of the Thai Cretaceous rocks have not yet been mentioned before, a preliminary result from recent studies on this aspect is included. These studies provide a basis for understanding paleoclimatology of the Cretaceous and this enables the Cretaceous history to be investigated, at least, in a preliminary way.

2. OUTLINE OF THE CRETACEOUS SYSTEM

2.1. Cretaceous distribution

In the Khorat Plateau, the nonmarine rocks are widespread throughout the area, consisting of 10 formations, namely (in ascending order): Huai Hin Lat, Nam Phong, Phu Kradung, Phra Wihan, Sao Khua, Phu Phan, Khok Kruat, Maha Sarakham, Phu Thok and Nachuak Formations. Of these, the Nam Phong, Phu Kradung, Phra Wihan, Sao Khua, Phu Phan, and Khok Kruat Formations belong to the Khorat Group. The Phra Wihan, Sao Khua, Phu Phan, Khok Kruat and Maha Sarakham Formations are believed to be the Cretaceous in age. The distribution of these formations shown in Figure 1 is described briefly below.

The Phra Wihan Formation, the lowermost formation of the Cretaceous, is broadly distributed along the western, eastern and southern parts of the Khorat Plateau. It also occurs in the northern and middle parts of the Phu Phan Range along the NW-SE trending anticlinal axes, which run from the north through the middle part, and less extensive in the southern portion.

The Sao Khua Formation is found throughout the western, eastern and southern parts of the Plateau and most parts of the Phu Phan Range along the NW-SE trending anticlinal axes.

The Phu Phan rocks are well exposed in most parts of the Phu Phan Range along the NW-SE trending anticlinal axes, particularly along the Mae Khong River banks. The rocks are formed as escarpment slopes, which delineate the outer rims of the Range.

The Khok Kruat Formation is assigned to the top of the Khorat Group, because the overlying Maha Sarakham Formation is separated from the Khok Kruat Formation by an unconformity [2, 3]. Both formations are well distributed in the central low-lying areas of the plateau and the outer parts of the Phu Phan Range bounded along the outer rims of the Phu Phan Formation with apparently conformable contacts. The sharp contact with the basal anhydrite of the overlying Maha Sarakham Formation is observed [4, 5] and is reported on seismic profiles [2].

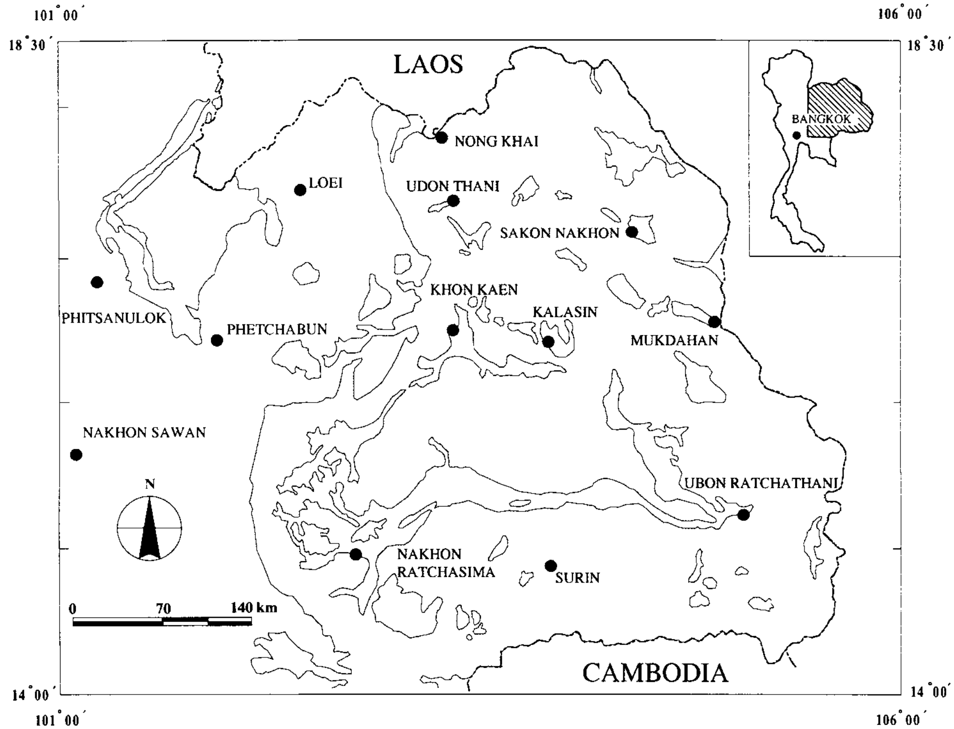
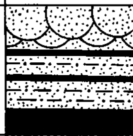
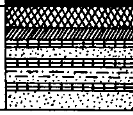


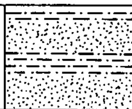
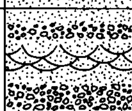
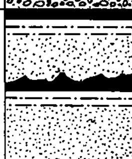
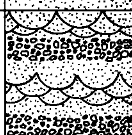
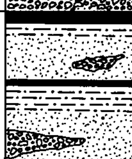
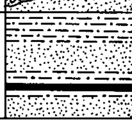
Figure 1. Map showing Cretaceous distributions (shaded) in the Khorat Plateau, northeastern Thailand. Inset is an index map showing the location of the study area.

2.2. Regional geologic settings

The Khorat Plateau in the northeastern Thailand is underlain partly by the Huai Hin Lat Formation deposited in a series of half-grabens formed due to the collapse of an overthickened crust during the Indosinian Orogeny, which was initiated by the collision of the Shan Thai and Indochina Plates [6]. Meanwhile the granite intrusion and andesitic volcanism to the east and west of the collision zone occurred [7]. The zone is marked by the Loei-Petchabun Fold Belt on the western rim of the Khorat Plateau. The sinistral NW-SE and dextral NE-SW conjugate strike-slip faults from across the two colliding plates [8] resulting in a series of nonmarine half-graben basins. Subsequent compression in the Late Triassic terminated this extension and was followed by widespread deposition of Jurassic-Cretaceous continental sediments of the Khorat Group, which filled up the Khorat Basin that formed as a result of thermal subsidence [3]. These Khorat Basins were partially inverted during the Late Cretaceous-Early Tertiary producing large wave-length structures around the margin of the Khorat Plateau. During this time the Phu Phan anticlinorium was formed through the inversion of the underlying Permian

Group	Formation	Lithologic column	Description	Thick - ness (m)	Age
	Phu Thok		Red sandstone, siltstone, claystone	100-550	Late Cretaceous - ?Early Tertiary
	Maha Sarakham		Brick-red siltstone and sandstone, thick beds of salt, gypsum and anhydrite	600-1,000	Late Cretaceous

UNCONFORMITY

KHORAT	Khok Kruat		Sequences of reddish-brown sandstone and siltstone with fossils of vertebrates	430-700	Early Crertaceous
	Phu Phan		Sequences of grayish-white, sandstone and conglomeratic sandstone with plant remains	80-140	Early Cretaceous
	Sao Khua		Sequences of reddish-brown sandstone and siltstone and claystone; fossils of vertebrates, bivalves and palynomorphs	200-720	Early Cretaceous
	Phra Wihan		A sequence of whitish-gray sandstone and conglomeratic sandstone; fossils of dinosaur tracks and palynomorphs	50-140	Early Cretaceous
	Phu Kradung		Sequences of maroon claystone, siltstone, sandstone and occasional conglomerate; fossils of vertebrates, bivalves and palynomorphs	800-1,100	Jurassic
	Nam Phong		Reddish-brown sandstone, siltstone and claystone; fossils of dinosaurs	100-1,500	Late Triassic

UNCONFORMITY

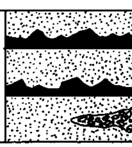
	Huai Hin Lat		Sequences of fluvio-lacustrine deposits: dark-gray mudstone, sandstone and occasional conglomerates	100-400	Late Triassic

Figure 2. Lithologic column of the Cretaceous and related rocks in northeastern Thailand.

and Triassic basins during this compressional event, which was probably caused by the collision of Burma with Shan-Thai [9] due to the arrival of the Indian Plate. The inversion created a rimmed basin similar to the present day topography of the Khorat Plateau and resulted in the development of two restricted basins (Udonthani-Sakon Nakhon and Khorat-Ubon Ratchathani Basins), within which the evaporites of the Maha Sarakham Formation were deposited. Mouret et al. [10] have calculated that up to 3000 m of the post-Phra Wihan sediments were eroded during the Late Cretaceous-Early Tertiary period..

3. PHYSICAL ENVIRONMENTS

As mentioned earlier, the Phra Wihan, Sao Khua, Phu Phan, Khok Kruat and Maha Sarakham Formations are believed to be the Cretaceous in age. The lithologic column of the Cretaceous and related rocks is shown in Figure 2. Over 2500 m of Cretaceous rocks are reported in the Khorat Plateau. The physical environments of these formations are interpreted as having been deposited under nonmarine conditions. Generally, the Sao Khua and Khok Kruat Formations are composed mostly of reddish-brown, fine- to medium-grained sandstones interbedded with siltstones and claystones with occasional conglomerates, caliches, calcrete nodules, and bedded-nodular silcretes, the intervening Phra Wihan and Phu Phan Formations consist of light-gray, fine- to coarse-grained quartzitic and conglomeratic sandstones with rare, gray siltstones and mudstones. The overlying Maha Sarakham Formation consists of red to reddish-brown clastic rocks with disseminated salts and gypsum. The physical changes observed from these formations, particularly at the base of each formation, are composed mostly of the conformable contacts, which can be clearly seen as sequence boundaries due to changes in depositional environments from the meandering river to braided river systems (e.g. from the Phu Kradung to Phra Wihan Formations or from the Sao Khua to Phu Phan Formations) or vice versa except for the boundary between the Khok Kruat and Maha Sarakham Formations, which have been interpreted as unconformable contacts [2, 3]. The temporal and spatial changes of Cretaceous rocks in the Khorat Plateau are described briefly as follows:

The meandering river deposition of the Middle-Upper Jurassic Phu Kradung Formation, became subject to tectonic uplift around the Khorat Basin. Significant amounts of clastic materials were transported by braided rivers to the basins, forming the Lower Cretaceous Phra Wihan quartzitic sandstones. Paleocurrent directions measured from cross-beddings in many outcrops suggest that the sediments were transported from various sources, but mostly from the northwestern parts of the plateau.

The Sao Khua Formation starts with rocks that consist of an alternation of reddish-brown silty claystones, siltstones and fine- to medium-grained sandstones with numerous thin caliches, carbonate nodules, and bedded and nodular silcretes. These rock types are different from those of the underlying Phra Wihan quartzitic sandstones. This difference is due to the gradual change in depositional environments from the braided to meandering systems. In places, the formation shows a sequence of fine-grained sandstones interbedded with reddish-brown siltstones with thinning-upward sequence and slump units grading up into gray, laminated siltstones rich in trace fossils. The formation is conformably overlain by the Phu Phan Formation consisting of light-gray, conglomeratic sandstones with subordinate conglomerates. The bivalves fragments are abundant along the base of the Phu Phan sandstones. Such gradual change can be seen locally in the eastern part of the Plateau. Paleocurrent directions measured from various localities for the Phu Phan Formation indicate that the paleocurrents were running towards the west and southwest except for a few places in the northern part of the Phu Phan Range, where the directions are variable due to the local uplifting in the western rim of the Range [10].

The Khok Kruat Formation is assigned to the top of the Khorat Group, because the overlying Maha Sarakham Formation is separated from the Khok Kruat Formation by an unconformity [2, 3]. This unconformity can be observed from well-log analyses, in which the two formations have the difference in velocity. The formation is conformably underlain by the Phu Phan Formation. The sharp contact with the basal anhydrite of the overlying Maha Sarakham Formation is observed [4, 5] and is reported on seismic profiles [2]. In general, the formation consists of reddish brown, fine- to medium-grained sandstones, siltstones and mudstones; conglomerates are also present. The formation was deposited by meandering rivers. Total thickness of this formation is unknown.

In general, the Maha Sarakham Formation is exposed in the low-lying areas particularly in the central part of the plateau. The uppermost parts of the formation, where the formation can be seen, the rocks consist of red to reddish-brown, fine-grained, laminated and small-scale cross-bedded sandstones interbedded with siltstones and mudstones with disseminated salts and gypsum.

Over 3000 m of sediments that were deposited in the Khorat Basin during the Cretaceous may have been reflected with sea-level changes. It is well known that during the Cretaceous long term sea levels reached extremely high stands [11, 12, 13], in particular, at the middle Cretaceous [14]. Based on sequence stratigraphy, large amounts of clastic materials should be supplied from widely exposed land areas during low stands of sea level, vice versa during high sea levels. However, high rates of clastic sedimentation do not always occur during sea-level low stand, but also reflect tectonic uplift and highly active magmatism, which accelerates weathering and erosion that increase the supply of large amounts of clastics to sedimentary basins, for example, during the peak stages of the global Cretaceous transgression in Japan [15].

4. BIOLOGICAL ENVIRONMENTS

Mesozoic nonmarine vertebrate and invertebrate faunas, and floras from the Khorat Plateau range in age from Late Triassic to early Late Cretaceous. The Cretceous biota of the Khorat Group and the Maha Sarakham Formation in the Khorat Plateau, northeastern Thailand, is summarised in Table 1. Biological changes can be observed in both space and time, which will be corresponding to the changes in physical environments.

Before the Cretaceous, the Jurassic vertebrate and invertebrate faunas are found in the Phu Kradung Formation. They include the crocodilian *Sunosuchus thailandicus* [16, 17]; sauropod and theropod dinosaur remains, which are more reminiscent of those of the euheropodid dinosaurs from the Jurassic of China [18], such as *Euhelopus* [19] or *Mamenchisaurus* [20]; vertebral elements of temnospondyl amphibians [21]; the bivalves *Protomida thailandica* (Hayami) and palynomorphs *Cyathidites minor* [3, 22]. These biotas are usually found in channel conglomerates and floodplain sandstones and siltstones of meandering river deposits. In places, silicified gymnospermous woods of possibly *Pinus* sp. are found embedded in maroon claystones. This wood is well-preserved in both trunks and branches; no growth rings are present. Wood is more abundant than in the Sao Khua and Khok Kruat Formations.

After changing in physical environments from the meandering to mainly braided river system, from the Jurassic Phu Kradung to the Cretaceous Phra Wihan Formations, the faunas

Table 1
Summary of the Jurassic-Cretaceous biota of the Khorat Group in the Khorat Plateau, northeastern Thailand

	AGE	FORMATION	VERTEBRATE	INVERTEBRATE	PALYNOFLORA
	Cenomanian-Albian	Maha Sarakham			
	Albian-Abtian	Khok Kruat	Dinosaur: *Psittacosaurus sattayaraki* *Thaiodus ruchae*		
Cenozoic	Abtian-? Barremian	Phu Phan	Dinosaur: *Irenesauripus*		*Corollina* sp. *Cyathidites minor* ?*Todisporites* sp.
		Sao Khua	Dinosaur: *Phuwiangosarus sirindhonae* *Siamotyranus isanensis*	Bivalves: *Trigonioides* sp. *Plicatounio* sp. *Trigonioides* (s.s) *trigonus* *Trigonioides*? cf. *guangxiensis* *Unio* sp., *Koreanaia* sp.	*Vitreisporites* sp. *Phedripites* sp. *Cyathidites* sp. *Cyathidites minor* ?*Araucariacites australis*
	Barremian-Berriasian	Phra Wihan			*Dicheiropollis etruscus*, *Corollinao* sp., *Araucariacites* australis, *Ischyosporites* cf. *variegatus*, *Cleichenidites senonicus*, *Laevigatosporites* sp., *Perinopollenites elatoides*, *Callialasporites dampieri*, *Anaplanisporites dawsonensis*, *Apiculatisporites* sp., *Osmundacites wellmanii*, *Todisporites minor*, *Kraeuseliporites* sp., *Concavissmisporites* sp., *Cicatricosisporites augustus*.
Jurassic	Late	Phu Kradung	Crocodilian: *Sunosuchus thailandicus*		

and floras are less abundant and diverse. Although many theropod dinosaur footprints are found in the Phra Wihan Formation at Phu Wiang [23] and Phu Faek [24], most of which cannot be identified in detail and have no clear biogeographical implications [18]. A Middle Jurassic age is given for the formation on the basis of plant fossils and arthropods [25], but this interpretation does not correspond to ages and stratigraphic relations with the overlying Sao Khua and underlying Phu Kradung Formations. The Phra Wihan Formation is, therefore, dated as Early Cretaceous (Berriasian to Berremian), which is supported by palynomorph data [3, 22].

The Sao Khua Formation, which overlies the Phra Wihan Formation, has yielded the richest and most diverse vertebrate fauna hitherto found in Thailand. It contains fresh-water hybodont sharks, actinopterygian fishes, turtles, crocodilians, and dinosaurs [18]. The crocodilian *Goniopholis phuwiangensis* [26] is described together with the theropod *Siamotyranus isanensis* which is considered as the oldest and most primitive known tyrannosaurid [27], and the sauropod *Phuwiangosaurus sirindhornae* [28]. Other dinosaurs include a new ornithomimosaurs [24]. Apart from vertebrate fossils, the Sao Khua Formation also contains the bivalves *Plicatounio*, *Unio*, *Koreanaia* and *Trigoniodes* [29], *Trigonioides* (s.s.) *trigonus*, *Trigonioides ?* cf. *gaungsiensis* (Sha Jin-jeng, personal comm.). Pieces of silicified wood in reddish-brown claystones are occasionally found in a few exposures. According to Racey et al. [3, 22] and Hahn [30], abundant palynomorphs have been described and the Cretaceous age is given for the formation. This age corresponds to that on the basis of dinosaurs and bivalves, in which the former is probably Valanginian to Barremian [27] and Early Cretaceous for the latter [29].

Biological changes are observed from the Sao Khua to Phu Phan Formations, and the decrease in diversity is very similar to that from the Phu Kradung to Phra Wihan Formations. Only dinosaur footprints are observed in sandstone beds at Phu Luang, in the Phu Luang Wildlife Sanctuary, near Loei [31]. Some palynomorphs from the Phu Phan Formation are also described e.g. *Cyathidites minor* [3, 22].

As mentioned earlier, the physical changes from the Phu Phan to Khok Kruat Formations are from the braided river system to the meandering river system, respectively. Thus the Khok Kruat Formation has yielded remains of various vertebrates, which can be compared with forms from other parts of the world [18]. Age determinations given for the formation are dated with relative precision. The occurrence of the fresh-water shark *Thaiodus ruchae* in the Khok Kruat Formation and also in the Takena Formation of the Lhasa block of Tibet, which is dated as Aptian-Albian on the basis of foraminifera [32], thus suggesting a similar age for the former formation. The vertebrate faunas from the Khok Kruat Formation includes the fresh-water shark *Thaiodus ruchae*, ceratopsian *Psittacosaurus sattayaraki* [33], and the remains of turtles, crocodilians and theropod dinosaurs. Pieces of silicified wood in reddish-brown claystones and siltstones are rare.

The Cretaceous Maha Sarakham Formation is interpreted as having been deposited in evaporitic conditions, so that fossils are usually rare. Based on palynomorphs the Albian-Cenomanian age is given for the formation [2]. This age corresponds to the Aptian-Albian age of the underlying Khok Kruat Formation mentioned above. Due to the increase in aridity from the Maha Sarakham to the overlying Phu Thok Formation, vertebrate faunas are very rare during this period.

5. CLIMATOLOGICAL ENVIRONMENTS

Most, if not all, of the nonmarine formations in the Khorat Plateau preliminarily examined so far contains paleosols and are found at the topmost part of each cycle, e.g. paleosols in the Sao Khua Formation showing calcrete horizons overlain by channel sandstones (Figure 3a) and mudcracks can be seen on the top of the horizons (Figure 3b). Thus the paleosols found in these formations can be used as paleoclimate indicators, because they provide a potentially useful information, under which it formed [34]. Paleoclimatological interpretation based on paleosols are becoming increasingly common (e.g. Mack [35]), but some studies are confined to the K-T boundary (e.g. [36, 37, 38, 39, 40]). Paleosols found in the Thai Cretaceous rocks are now in preliminary stages of investigation, but some field data obtained from the Jurassic Phu Kradung and Cretaceous Sao Khua and Khok Kruat Formations may be used as one of the indicators of paleoclimates.

Apart from paleosols found in the Jurassic Phu Kradung Formation and the Cretaceous Sao Khua and Khok Kruat Formations, nodular and bedded silcretes can also be found in the Sao Khua Formation, and are more abundant in the Sao Khua Formation than in other formations.

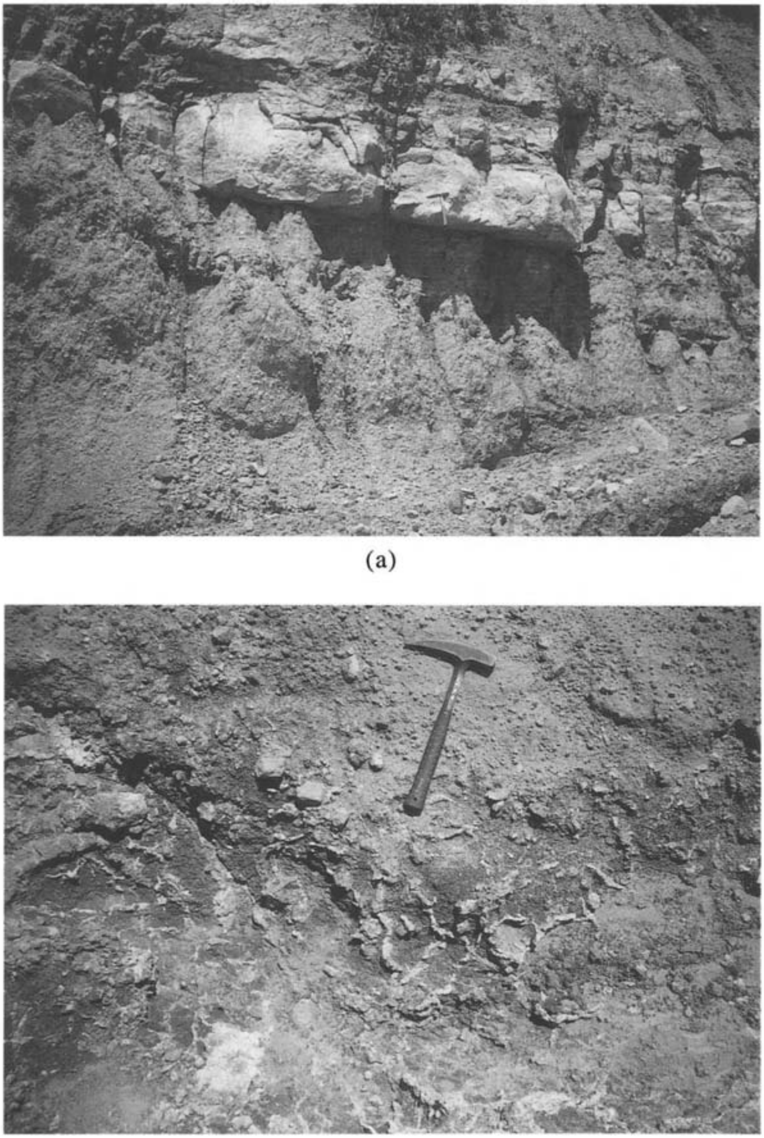
5.1. Descriptions of paleosol and silcrete

5.1.1. Recognition of paleosol horizons

Previously known calcrete horizons found in reddish-brown claystones of the Sao Khua Formation are interpreted as a result of recent investigations, to have been deposited *in situ* by the meandering fluvial system; the term "paleosol" will be used for these horizons in this paper. Paleosols in the Sao Khua Formation are found throughout the Khorat Plateau, but are more abundant in the eastern part. These paleosols can be classified into two main types: calcic and non-calcic. Whereas the calcic paleosols are predominated in meandering fluvial sequences, the non-calcic varieties are usually found in the braided fluvial sequences. The calcic paleosols of the Sao Khua Formation, in particular, range in thickness from 0.5-5 m (average 2 meters) and consist of horizons with calcrete nodules ranging in size from 0.3 x 0.6 - 2.0 x 5.0 cm and reddish-brown, non-oriented clay matrix. These horizons are interpreted as the C-horizon of soil profile, because root traces and bioturbations are rare. In general, paleosols are found at the top of the fining- and thinning-upward sequences and overlain by channel sandstones and on the top along bedding planes, boxwork-like calcareous mudcracks can be seen. In some places, these types of paleosols were eroded and calcrete nodules were re-deposited as channel conglomerates on the top. Thin weathering profiles on the top of these paleosols are common. Vertebrate and invertebrate fragments are absent in these paleosols, but can be usually found in conglomerates.

5.1.2. Silcretes in the Sao Khua Formation

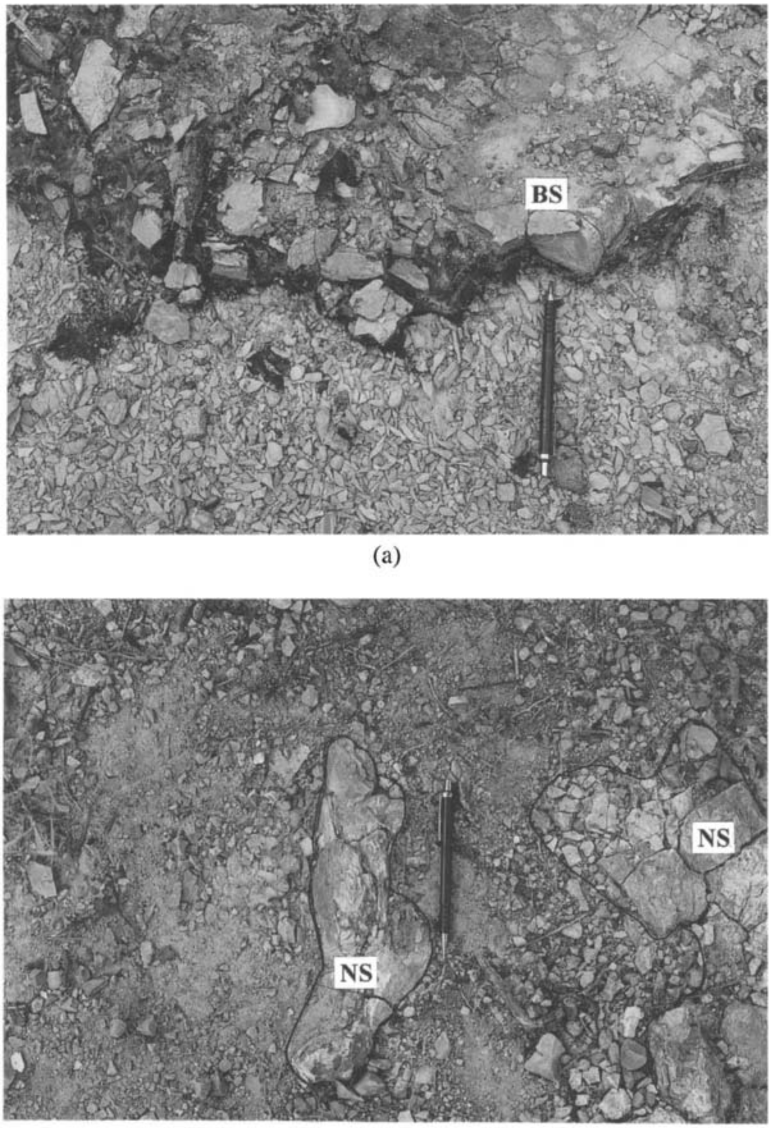
Silicified wood-like horizons in the Sao Khua Formation are interpreted as silcretes that have been formed by weathering products of silica-rich parent rocks in low-lying paleo-topographic areas. Silcretes found in the Phu Sra Dok Bua National Park as bedded (5-10 cm thick) forms (Figure 4a) and nodular forms (10×25 cm) (Figure 4b) mostly in the Sao Khua Formation, may provide information about ancient terrestrial environments. The silcretes interbedded in this formation are locally silicified layers at almost the top of the fining-upward units, in which the overlying calcrete horizons are present. In some horizons, slickensides and boxworks in these silcretes are found due to shrinkage and swelling of clay minerals, indicative of temporary subaerial exposure of the sediments and later silicification.



(a)

(b)

Figure 3. (a) Photograph of paleosols as a calcrete horizon underlying crevasse splay sandstones in the Sao Khua Formation (Hammer for scale is 28 cm long).
(b) Photograph of boxwork-like mudcracks on the top of paleosols in the Sao Khua Formation (Hammer for scale is 28 cm long).



(a)

(b)

Figure 4. (a) Photograph of bedded silcretes (BS) overlying grayish-brown claystones of the Sao Khua Formation (Pencil for scale is 14 cm long).
(b) Photograph of nodular silcretes (NS) embedded in grayish-brown claystones of the Sao Khua Formation (Pencil for scale is 14 cm long).

This type of silcretes is very similar to those found in continental Cenozoic of western Portugal and is interpreted as having been deposited in local swampy environments [41].

5.1.3. Paleoclimatic significance

In general, Cretaceous paleoclimate is interpreted to have been warmer and more equable than modern climate conditions attributed to a combination of high sea level, a large area of land in the subtropics, lack of continental glaciers, and high level of atmospheric CO^2 (see Barron [42] and Hallam [43]).

The principal data used for supporting a warm, equable Cretaceous are temperature estimates of seawater based on $\delta^{18}O$ values of molluscs and foraminifera (see Savin [44]) and the global distribution of plant fossils [45, 46, 47]. However, the reassessment of Cretaceous paleoclimate arises from new isotopic and paleobotanic data (e.g. [48, 49, 50]), the presence of ice-rafted boulders suggestive of polar ice caps [51], and numerical modeling that cannot always account for the presumed warmth of the Cretaceous by applying standard climatic variables to Cretaceous paleogeography [52, 53, 54, 55, 56, 57]. These new data underscore the fact that Cretaceous paleoclimate is far from completely understood, and they emphasize the need for more data or perhaps new approaches to paleoclimate interpretation.

In order to use paleosols as paleoclimatic indicators, it is necessary to establish that differences among paleosols are not the result of other factors, such as parent material, slope, age, and vegetation [58]. Actually, the paleosols found in the Sao Khua Formation are similar to those found in the Phra Wihan Formation, but the only difference is that the former is calcic but the latter is non-calcic paleosols. The calcic paleosols are usually found in arid to semi-arid environments. The grain sizes and composition of parent material in both series are comparable, in which the paleosols have been developed within claystones and siltstones. In case of slope, upon which the paleosols developed, the paleosols in both series developed either within the floodplain or on the top of abandoned channels, so the slopes were probably quite low and varied little with time [59, 60, 61]. Also the age or degree of maturity factor is controlled by considering those paleosols developed in more or less the same horizon and depositional cycle. The soil-vegetation relationship is the most difficult to determine for the Cretaceous paleosols of the Khorat Plateau, because there have been no detailed paleobotanical studies of the formations. However, conifers collected from the Jurassic Phu Kradung Formation and a few samples of unidentified wood in the Cretaceous Sao Khua and Khok Kruat Formations, are the known data for fossil plants. Apparently, silicified woods collected from the Khok Kruat Formation are angiosperms, which had already appeared by the Aptian time [36]. It seems unlikely that vegetation changes alone could have accounted for differences between each paleosol in the Khorat Plateau. In the paleosols in the Sao Khua Formation and related rocks, calcrete horizons, mudcracks, and thin weathering profiles of these paleosols are indicative of, at least, semi-arid environments during the formation.

Silcretes found in the Sao Khua Formation and related rocks may also be used as paleoclimate indicators, but it is still questionable because of uncertainty with respect to the climatic significance of relatively recent (Cenozoic) silcretes and the lack of recognition in older rocks. Although most authors tend to favor an arid to semi-arid climate for silcrete genesis [63], others have suggested that silcrete can also form in a humid environment [43, 64]. Recent work in southern Africa has led to the development of geochemical and petrographic criteria for discriminating silcrete formed in arid and humid paleoclimates [65, 66]. However, silcretes found in the Sao Khua Formation are compared to those of South Africa and Canada in terms of petrography (Table 2) and geochemistry. This comparison led to the preliminary understanding that the silcretes in the Sao Khua Formation show several charac-

Table 2
Comparison of Cenozoic arid and humid climate silcretes from southern Africa (after Summerfield [65]) and Proterozoic arid climate examples from Canada (Ross and Chiarenzelli [67]) with Thai Cretaceous semi-arid specimens

	Cenozoic examples (Summerfield [42])		**Proterozoic examples** (Ross and Chiarenzelli [46])		**Cretaceous examples** (This study)
	Humid	Arid	Thelon Basin (Arid)	Hornby Bay Group (Arid)	NE Thailand (Semi-arid)
Petrographic Features					
Optically continuous overgrowths	A	U	U	U	U
Chalcedonic overgrowths	A	U	U	U	U
Length-slow calcedonic vugfills	A	C	C	C	C
Glaebules					
Ti-rich	C	A	A	A	A
Fe-rich	C	U	U	U	not available
Colloform features	C	A	A	A	A
Silicified clays	C	C	C	A	U
Associated Features					
Deep-weathering Profiles	C	A	C	A	A
Thin-weathering Profiles	A	C	A	C	C
Eolianites	A	C	C	C	U
Evaporites	A	C	C	C	U
Calcrete	A	C	A	A	U
Playa sediments	A	C	C	U	not available

A = Absent; U = uncommon; C = Common

teristics, which suggest development during semi-arid climatic conditions. Semi-arid climate lithogenesis of these silcretes is suggested by the void fills composed of length-slow fibrous chalcedony and microcryptalline quartz. Ti-rich glaebular and colloform structures indicative of humid environment silcretes are not seen in the Thai silcretes. Thin weathering profiles found in both paleosols and silcretes also support arid to semi-arid environments. Geochemically, this is further supported by the low concentration of titanium dioxide ($TiO^2 \leq 0.03$) (see Table 3) in silcretes, in which the humid environment has high concentration of Ti (> 2.0% of TiO^2). Chemical analyses shown in Table 3 indicate that the Thai silcretes are quite different from those of Australia, which have been interpreted as having been formed in wetter climate [68].

Table 3
Comparison of a few major element analyses for silcretes collected from the Sao Khua Formation, northeastern Thailand with Victorian and South Australian silcretes. Source: Thai silcretes analysed by the Analytical Division, Department of Mineral Resources, Bangkok; Australian silcretes (modified after Webb and Golding [68])

Locality	Lithology	Sample No.	SiO_2	Al_2O_3	TiO_2
NE Thailand	Bedded silcrete	576/41	91.89	0.79	nil
	Nodular silcrete	576/41	86.24	2.97	0.03
Victoria (eastern association)	Silcrete	8580/31	98.60	0.42	0.07
	Silcrete	8580/73	98.82	0.02	0.17
	Silcrete (median)	8580/70A,B	95.86	0.62	0.47
	Silcrete	8580/140A	95.65	1.05	0.88
	Silcrete (median)	8316/24,27	98.82	0.26	0.15
South Australia (inland association)	Pedogenic silcrete (top of horizon)	8580/107B	98.61	0.06	0.26
	Pedogenic silcrete (base of horizon)	8580/107C	96.65	0.10	0.26
	Pedogenic silcrete	8580/11A	92.48	1.59	0.24
	Groundwater silcrete	8580/106A	96.99	0.23	0.05
	Groundwater silcrete	8516/67	97.14	0.35	0.26

6. TOTAL IMPACTS ON ENVIRONMENTAL CHANGES

After the collision of the Shan-Thai and Indochina Plates in the Late Triassic, the Huai Hin Lat Formation was deposited in a series of half-grabens formed due to the collapse of an over thickened crust during the Late Triassic Indosinian Orogeny. This was followed by widespead deposition of Jurassic-Cretaceous sediments of the Khorat Group, which filled up the Khorat Basin that sagged due to thermal subsidence [3]. Environments of deposition, biological environment and paleoclimate play the important roles in controlling the deposition of more than 2500 m thick of the Cretaceous and related rocks in the Khorat Plateau. From Late Triassic to Cretaceous, the sediments are interpreted as having been deposited in nonmarine conditions (meandering and braided river systems) during semi-arid paleoclimate with occasional slightly humid and eventually becoming evaporitic condition in arid paleoclimate at the Late Cretaceous.

By the end of Triassic, the nonmarine sediments represented by the Nam Phong Formation were pervasively deposited in the Khorat Basin by meandering rivers in semi-arid paleoclimate. Only dinosaur bones have been found in the strata of this period.

In the Jurassic, long period of meandering river deposits with semi-arid condition still existed resulting in the deposition of more than 1000 m thick nonmarine red beds. Repetitive cycles of meandering channel and floodplain deposits can be clearly seen in various outcrops together with local pedogenesis on the top of these floodplain sequences. Invertebrate and vertebrate faunas are more abundant and diverse than those in the Triassic. Bivalves, crocodilian, turtles, fishes, dinosaur bones and petrified woods have been found in many sections.

In the Early Cretaceous (Barremian-Berriasian), the depositional environments changed from the meandering river system to the braided river system due to the changes from semi-arid to slightly humid climate as demonstrated by the presence of some plant remains in mudstones and the uplifting in surrounding areas of the plateau. Accordingly, huge volumes of sediment were redeposited into the basin giving rise to the deposition of thick-bedded quartzitic sandstones and occasional mudstones of the Phra Wihan Formation. During this time, faunal abundance and diversity slightly decreased. Only dinosaur tracks have been recorded on the sandstone beds and some palynomorphs in mudstones.

During Aptian-?Barremian, the prevailing braided river system has changed to meandering river together with the changes from slightly humid to a semi-arid paleoclimate. Repetitive cycles of meandering channel and floodplain deposits can also be clearly seen in various outcrops together with local pedogenesis on the top of these floodplain sequences of the Sao Khua Formation, and are more conspicuous than from the Jurassic. Invertebrate and vertebrate faunas are rich and diverse, containing fresh-water hybodont sharks, actinopterygian, fishes, turtles, crocodilians, theropod and sauropod dinosaurs, bivalves and palynomorphs. The paleoclimate during this time was more semi-arid than that in the Late Triassic-Jurassic, as can be seen from various stratigraphic sections, which contain many calcrete horizons and some bedded and nodule silcrete layers. After a prolonged meandering river system of the Sao Khua Formation, another large amounts of sands and gravel from the northeastern part of the plateau were also laid down in the the Khorat Basin by high-energy braided rivers. Paleocurrent analyses from conglomeratic sandstones of the Phu Phan Formation have confirmed the paleocurrent directions. Due to high-energy currents, invertebrate and vertebrate remains preserved in the siltstones and claystones of the Sao Khua Formation were removed and redeposited at the base of the Phu Phan conglomeratic sandstones. The paleoclimate during this time was humid and the eastern region has been subjected to strong tectonic uplift.

Following a period of deposition of high-energy braided rivers for the Phu Phan Formation, the paleoclimate changed from humid to semi-arid and eventually to arid until the end of the Cretaceous. Although depositional environments were confined to a meandering river system, evaporitic conditions prevailed at the Late Cretaceous. This is indicated by cyclic deposition of channel sandstones and flood-plain sequences of the Khok Kruat Formation followed by the deposition of three-layers of salt and evaporitic minerals of the Maha Sarakham Formation. Vertebrate remains and some bivalves are found locally in the former Formation, but are absent in the latter.

In conclusion, the physical, biological and climatological changes in the Cretaceous rocks of northeastern Thailand resulted in the repetitive changes from the meandering river system to the braided river system corresponding to paleoclimatological changes from semi-arid to slightly humid conditions, and finally arid conditions in the Late (possibly latest) Cretaceous. The abundance and diversity of faunas and floras affected by physical and climatic changes in both space and time. Continued studies in stratigraphy, paleontology, petrography, geochemistry and paleoclimatology of the Cretaceous sequences in northeastern Thailand should provide a basis for future testing and refining models of the Thai Cretaceous environmental evolution.

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Spatio-temporal patterns of environmental changes in Late Cretaceous sequences of Central India

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Late Cretaceous strata of Central India–the Bagh and Lameta Beds/Intertrappean Beds are patchily distributed as thin sedimentary sequences in the rift zones of Central India. These strata record the spatio-temporal variations of Late Cretaceous environments in an Indian lithosphere influenced by rifting, plume activity (Kerguelen and Reunion) and continental flood basalt volcanism. Analysis of depositional environments of these strata reveal that an epicontinental seaway advanced from the west during the Turonian (?) and inundated areas a few hundred kilometers inland in the Lower Narmada Basin. Earlier phases of rifting related sedimentation included estuarine facies (Nimar Group) in the Lower Narmada Basin and fluvial channel facies (Jabalpur Group) in the upper parts of the basin.

Fluvial/lacustrine/palustrine and sheetwash deposits constitute the infra- and intertrappean sequences which are broadly coveal with the Maastrichtian Deccan lava flows. Surface uplift, possibly plume related, caused withdrawal of the sea from the Lower Narmada Basin; and raised the land surfaces in the upper parts of the basin. This led to the occurrence of widespread subaerial exposure conditions as revealed by the regionally persistent dinosaur egg-shell-bearing Lower Limestone (Lameta Beds) – a palustrine-pedogenic facies.

In some regions of the central Indian lithosphere the Infratrappean Lameta Beds are made up of alluvial-lacustral facies. Such areas possibly represent those parts of the Indian lithosphere effected by surface uplift and fracturing/faulting. Faulting-related relief produced adequate topographic differentiation to maintain alluvial-lacustrine systems.

1. INTRODUCTION

Late Cretaceous strata of Central India including the Bagh, Lameta and Intertrappean beds have attracted attention for more than a century because of a) their rich and varied dinosaurian

Discussions with several colleagues, more particularly Dr. D.M. Mohabey of the Geological Survey of India, Nagpur, helped to clarify many of the palaeoenvironmental aspects of the Lameta Beds of Nand-Dongargaon areas. Dr. S.C. Bhatt, Department of Geology, University of Delhi, rendered invaluable assistance in preparation of the manuscript. I am also indebted to the Indian IGCP-350 program in many ways, more particularly for providing opportunities for field discussions on many occasions, and particularly for the field excursion to the Narmada region in December, 1997 led by Professor Surinder Kumar of the Lucknow University.

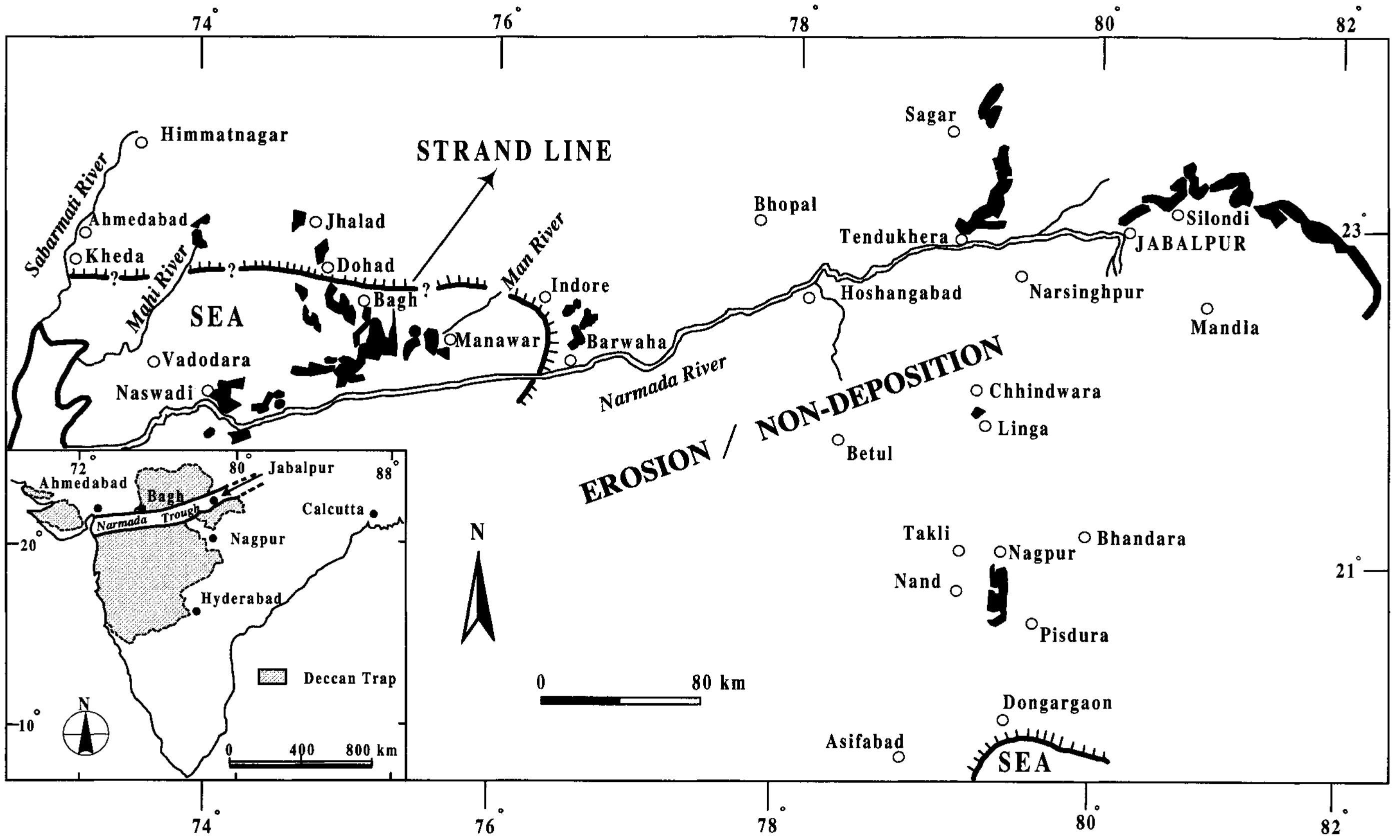

Figure 1. Location map of the Lameta and Bagh Beds of Central India with an index map of peninsular India showing the Lower Narmada region, west of Indore. The Bagh Beds are confined to the Narmada Trough region. Possible configuration of land and sea in the Maastrichtian is also plotted.

and other fossil materials, b) their association with a thick pile of lava flows–the Deccan Volcanics, and c) their importancce in the Cretaceous palaeogeographic reconstruction of India. Also, as a repository of some of the last dinosaurs in India [1], the Lameta Beds and Intertrappean strata constitute key sequences for understanding Maastrichtian palaeoenvironments and end-Cretaceous global changes [2, 3, 4, 5, 6]. Despite this aforementioned importance of Late Cretaceous strata of Central India, a synthesis of environmental changes/shifts of the region in space and time is not available.

The wide areal extent of the Late Cretaceous strata (Figure 1), occurring as disconnected patches spread over Central India, was emphasised by the early investigators [7]. Most previous work on palaeoenvironmental aspects, in contrast, focussed mainly on the Lameta Beds of the Jabalpur region [4, 5, 6, 8, 9, 10, 11]. Hence, regionally based palaeoenvironmental data of the widespread Late Cretaceous strata are meagre. Recent palaeoenvironmental work on the Lameta Beds of Maharashtra by Mohabey et al. [12] and Mohabey [13], on the Intertrappean Beds of Anjar (Kutch) by Khadkikar et al. [14], and on the Bagh Beds of the Lower Narmada region by Sood [15] permit an initial synthesis of spatio-temporal patterns of environmental shifts in the region.

Taken in conjunction with the extensive previous work on the Lameta Beds of the Jabalpur region, these data are used to understand spatio-temporal patterns of sedimentation in an area influenced by rifting, plume tectonics and related continental flood basalt volcanism. In such a tectono-sedimentary milieu, localised marine environments, changed over to predominantly subaerial exposure environments in the Lower Narmada region. Alluvial-limnic environments predominated in some other parts, for example, Maharashtra [12].

The purpose of this particular study is to synthesise the available data on palaeoenvironments of the Bagh Beds, Lameta Beds, and the Intertrappean sequences from Central India, and to evaluate the spatio-temporal patterns of environmental change/shift as a function of geodynamic factors.

2. SEQUENCES AND PALAEOENVIRONMENTS OF THE BAGH BEDS

The Bagh Beds occur mainly in the western segment of the >1200 km long Narmada depression (Figure 2), and consist dominantly of calcareous facies deposited in a Cretaceous embayment. Two persistent lithological units, the Nodular Limestone and Coralline Limestone occur in most sections. These carbonate sequences overlie the extensively occurring Nimar Group which contains cross-bedded sandstone, conglomeratic sandstones with subordinate clays and marls. The Nimar clastic facies are usually assigned to the Early Cretaceous, and represent continental to estuarine facies [16] deposited in an intracratonic rift-related basin – the Lower Narmada Basin which extends from Barwaha in the east to Naswadi in the west (Figure 2). The Nimar Sandstone directly overlies the Precambrian basement. Its thickness varies from 15 to 30 m in the east as against 150 m in the western parts of the basin [16]. These estuarine rocks are overlain by a transgressive facies – Nodular Limestone which is fossiliferous in its upper part, and contains bivalves, bryozoa, gastropods and ammonites. The overlying Coralline Limestone is a cross-bedded (-1m thick) calcarenite made up of skeletal detritus – mostly bryozoan fragments.

Relationships of the Bagh carbonate lithofacies with the overlying Lameta rocks – clastic and carbonate – are incompletely understood. Some workers [16, 17] recognise these as being

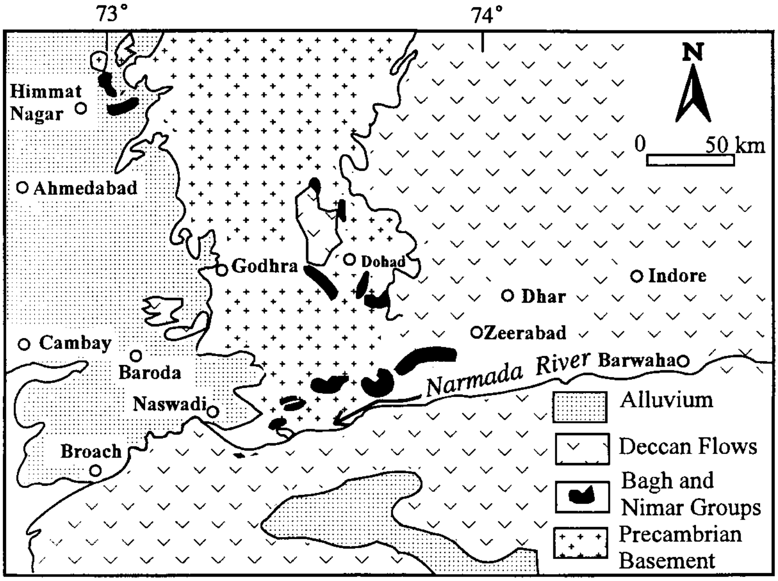

Figure 2. Map showing the distribution of the Bagh Beds in the Lower Narmada region and their relationship with the Deccan volcanics and the pre-Cambrian basement.

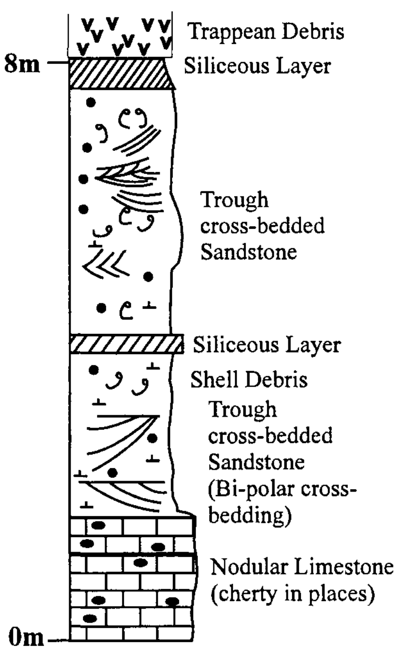

Figure 3. Litholog of the Bagh Beds (Nodular Limestone) and the supra-Bagh lithofacies at Phutlibaori.

conformable with the Bagh Beds whereas Roychaudhri and Sastry [18], based on detailed geological mapping, concluded that the Lameta Beds, where they are in contact with the Bagh Beds overlie them unconformably. New data [15, 17] reveal that in some sections (Figure 3) the Nodular Limestone/Coralline Limestone is overlain by a clastic facies. These green/white sandstones, up to 6 m thick, show bipolar cross-bedding and abundant shell debris. At Chakrud, oyster beds are associated with the white sandstones [17]. Petrographic data [15] of the cross-bedded calcareous green sandstone facies of the Phutlibaori section show that siliciclastic grains constitute about 50% of the grains with bioclastic grains making up about 25 to 30%. Felspar grains may occur up to 15% of the grain population.

More commonly, however, the Bagh carbonate lithofacies are overlain by a calcareous sandstone/calcrete containing complete dinosaur eggshells. Most workers [5, 12, 13, 19] favour a continental origin for them, particularly for the dinosaur eggshell-bearing lithology (variouly termed as limestone, calcareous sandstone, calcrete in the existing literature).

Analysis of the Late Cretaceous lithofacies of the Lower Narmada Basin (essentially to the west of Indore) permits the following palaeoenvironmental deductions:

a) deposition of the Nimar Sandstone in estuarine complexes consisting of channels separated by terrigenous mudflats and carbonate mudflats; sediment dispersal was mainly westwards [16];
b) deposition of the Nodular Limestone marking a transgressive facies system from the west;
c) deposition of the Coralline Limestone in a lowstand systems tract consisting of carbonate tidal complexes;
d) deposition of white/green cross-bedded shell-bearing sandstone representing local estuarine channel complexes in a tidal zone; and
e) deposition of dinosaur eggshell-bearing calcareous sandstones representing widespread subaerial exposure environments.

The Lower Narmada Basin was then influenced by the extensive Deccan volcanism which resulted in a thick cover of lava flows.

3. SEQUENCE AND PALAEOENVIRONMENTS OF THE LAMETA BEDS OF THE JABALPUR AND NAGPUR REGIONS

In contrast to the marine Bagh Beds (Turonian?) which are restricted to the Lower Narmada Basin, the infratrappean beds (Lameta Beds) are patchily distributed throughout the Narmada region. They are particularly well developed in the eastern part of the Narmada valley (for example, Jabalpur sub-region, Saugar sub-region) where they unconformably overlie granite basement, Precambrian metamorphic/sedimentary strata, or Gondwana strata. Despite some contrary opinions [9, 11, 20, 21], most recent studies [3, 4, 5, 6, 8, 12, 13] from the eastern and the central parts of the province concluded that the Lameta Beds are terrestrial deposits without any marine influence. Further, Tandon et al. [5] suggested that the Lameta Beds of the Jabalpur sub-region constitute a Maastrichtian regolith (Figure 4), exhumed by erosion of the covering Deccan Traps.

3.1. Jabalpur region

Detailed sedimentological descriptions of the Lameta sequences, including lithofacies analysis and palaeoenvironmental interpretations are available in recent work [4, 5, 6]. The basal Green Sandstone, 3-10 m thick, consists of green and white, medium- to coarse grained- and

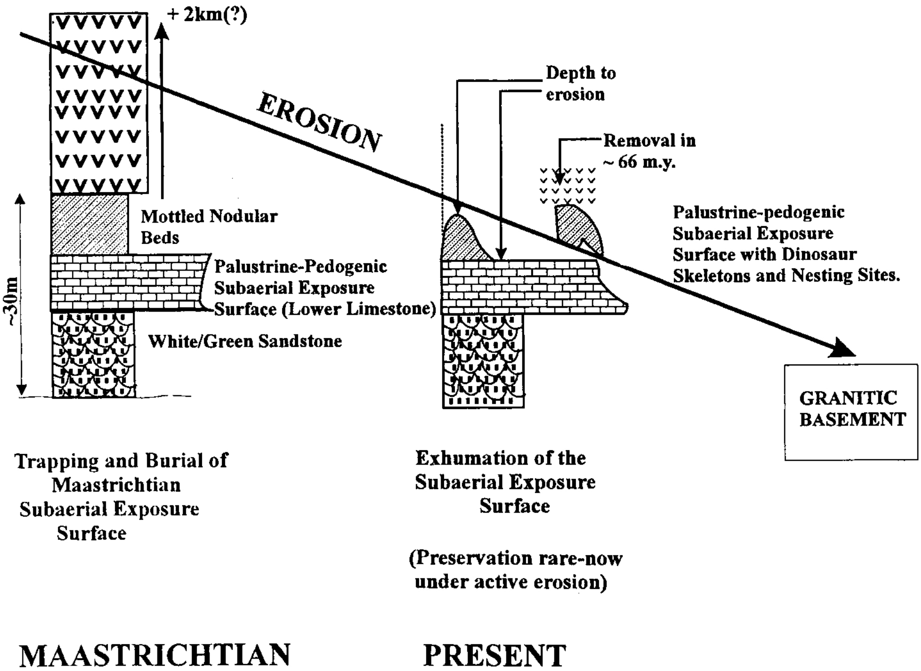

Figure 4. Stratigraphic model for the Lameta Beds of the Jabalpur sub-region; a Maastrichtian calcareous regolith currently under exhumation because of erosion of the Deccan volcanics.

pebbly, trough cross-stratified sandstones. Large, medium, and small-scale trough cross-bedding, channel scours, channel lags, and horizontal laminated muddy sandstones are common features. Individual co-sets of trough cross-beds measure up to 1 m; groups of co-sets are truncated by erosional surfaces. In places, upward fining units are noted. Also, parallel, laminated muddy sandstone, mud drapes and thinly inter-laminated mudstone and sandstone occur associated with the trough cross-bedded sandstones. Small mud clasts within the sandstones indicate contemporaneous erosion. On the basis of these features, the green sandstone is interpreted as two to three storied sand accumulations in shallow, braided, fluvial channels with stage fluctuations [5].

The Lower Limestone is 3-15 m thick in the Jabalpur sub-region. It contains the following facies: a) palustrine limestone, b) sandy, nodular, brecciated calcrete, c) calcareous grey siltstone with brecciated calcareous nodules, d) brecciated pisolitic calcrete, e) sandy and pebbly green marl, and f) calcrete nodule conglomerate. The most important attribute of the calcareous lithofacies of the Lower Limestone is *shrinkage*. This is manifested by the development of multiple generations of cracks, related brecciation fabrics, and micro-relief. Collectively, the characters of the Lower Limestone facies, and their lateral and thickness variations suggest formation in a sub-aerially exposed low gradient alkaline flat in an alluvial setting [5].

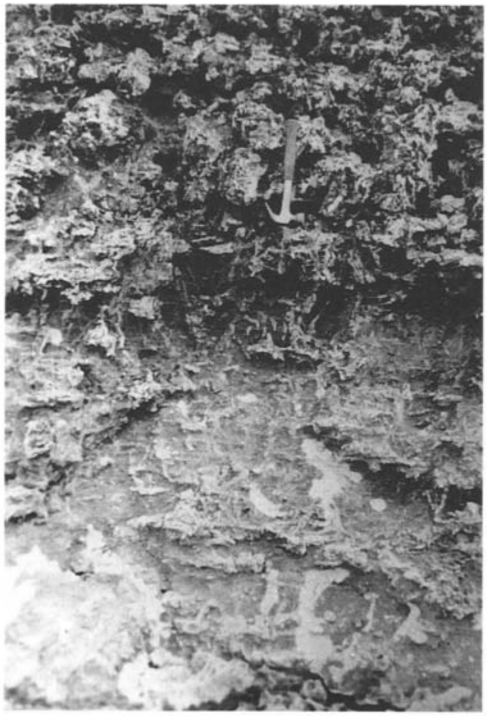

Figure 5. Field photograph of red silty and clayey very fine sandstone with abundant mottles and tubular structures overlain by a shrinkage related nodular calcrete. Location: Motttled Nodular Beds near Sivni village.

The Mottled Nodular Beds consist of red, silty and clayey, very fine-grained sandstone, sandy marl and calcrete (Figure 5). Important outcrop features include carbonate pipe structures, rhizocretions, green mottles, siliceous tubular structures and rarely laminar structure. From a detailed analysis of these strata in the Sivni section, Tandon et al.[6] showed that multiple calcrete profiles constitute the Mottled Nodular Beds. These profiles show clear variations in the distribution of carbonate morphologies (powdery, nodular and sheet calcretes). Profile development was governed mainly by the pattern and depth of shrinkage [6]. In the Sivni section, the Mottled Nodular Beds measure ~16.5 m, and are organised into a vertical sequence of fourteen profiles with variable characteristics. The calcrete/calcareous paleosols of the Mottled Nodular Beds are interpreted to have formed in the soil-vadose zone, usually within a few metres of the sediment surface.

Development of a vertical stack of successive calcrete profiles like those of the Sivni section requires discrete periods of high magnitude aggradation, large enough to bury the existent soil surface. The calcrete profiles of the Mottled Nodular Beds constitute a rare example of few widely spaced sheet flood sediment accretionary events being converted into a stack because of a favourable combination of climatic conditions, carbonate availability, and sediment starved conditions, related in some measure, to the contemporaneous Deccan volcanic events [6].

Figure 6a. Field photograph of palustrine limestone with abundant shrinkage features. Note the sandy calcrete in the fissure fills.

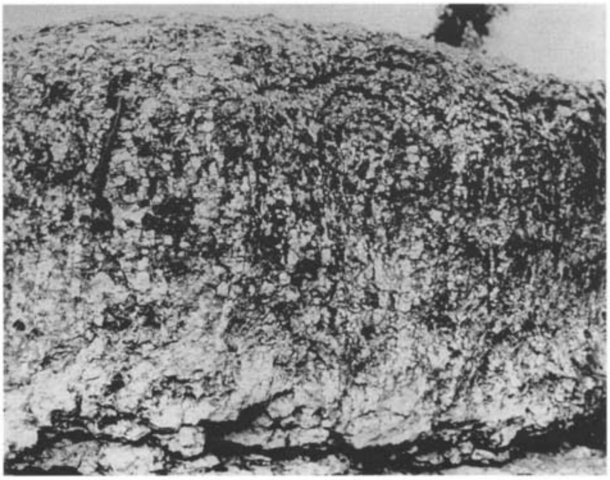

Figure 6b. Development of a strong vertical fabric due to shrinkage in the brecciated nodular calcrete of the Lower Limestone. Location: Lametaghat on the north bank of the Narmada river.

The Mottled Nodular Beds are developed locally in the Jabalpur sub-region, and are not recorded elsewhere in the Narmada region, where the Lower Limestone is the typical and persistent Lameta unit [22].

The Upper Calcified Sandstone consists of moderately sorted trough cross-bedded sandstone which is variably calcified [5]. The sandstones are quartzose, and in places, pebbly (jasper, chert, reworked siliceous tubular bodies from the underlying Mottled Nodular Beds) detritus is common. Green mudstone clasts are locally abundant. These deposits have been formed by widely occurring sheetflows/sheetfloods, mainly unchanneled [5].

The depositional model for the Lameta Beds of the Jabalpur sub-region show the following stages: a) fluvial braided channel marked by stage fluctuations and ephemerality, b) mixed carbonate clastic low relief subaerial plain in which carbonate deposition took place, mostly, in desiccating ephemeral lakes/ponds/palustrine flats, c) sheetwash redistribution in a near surface regolith with abundant calcareous pedogenesis, and d) unchannelised sandy sheetflows/sheetfloods.

Clearly the Maastrichtian Lameta Beds in the upper parts of the Narmada region of Central India developed predominantly under conditions of sub-aerial exposure.

3.1.1. Dinosaur eggshell-bearing limestone

The main persistent unit of the Lameta Beds, the Lower Limestone, occurs extensively in the entire Narmada valley and requires further consideration. Significantly, this is the dinosaur eggshell-bearing unit of the region [1]. As an illustrative example, facies elements of the Lower Limestone are shown in Figure 6. Palustrine limestone is associated with most other lithofacies types, for example LF1a calcretised fissure-fill sandstones are hosted in LF1 palustrine limestone; LF4 brecciated, nodular and pisolitic calcretes, and LF7 calcrete conglomerates contain large zones of disrupted palustrine limestone or fragments thereof. Lateral variations amongst the facies elements are very rapid, and are related to the variable degree of fissuring, collapse of fissures and sediment infill, and availability/re-distribution of sheetwash sediments.

The LF3 calcareous grey siltstone and LF6 sandy green marls contain an assemblage of fresh water aquatic pulmonate gastropods, charophytes and sauropod/theropod skeletal materials [23, 24]. These facies with aquatic biotas pinch out over a few tens of metres or grade laterally into subaerial exposure facies-calcretes and sheetwash deposits. These relationships indicate that the aquatic bodies were locally formed ephemeral carbonate ponds in a low relief subaerial plain effected to varying degrees by sheetwash sedimentation and calcareous pedogenesis.

Local variations in the regionally occurring dinosaur eggshell-bearing breciated (and nodular) sandy calcretes are significant. For instance, the degree and nature of sheetwash incorporated is variable to the extent, that in places (for example, Hathni section) it is referred to as a calcareous sandstone [17]. Also, the intensity of shrinkage characteristics of these calcretes is quite variable leading to variable degrees of intrastratal reworking. This has a bearing on the nature of the preservation of the sauropod eggshells (Figure 7). Commonly, more complete eggs are abundant in the Lower Limestone of the Lower Narmada region where the detritus is sandier and felspathic. Contrarily, the brecciated and fragmented sauropod eggshells occur invariably in the sandy, brecciated, nodular calcretes of the Lower Limestones of the Jabalpur sub-region. This clearly reflects shrinkage-related intrastratal reworking, possibly aided by the shrink-swell properties of volcanogenic clays (montmorillonite) derived from the contemporaneous lava flows [25, 26].

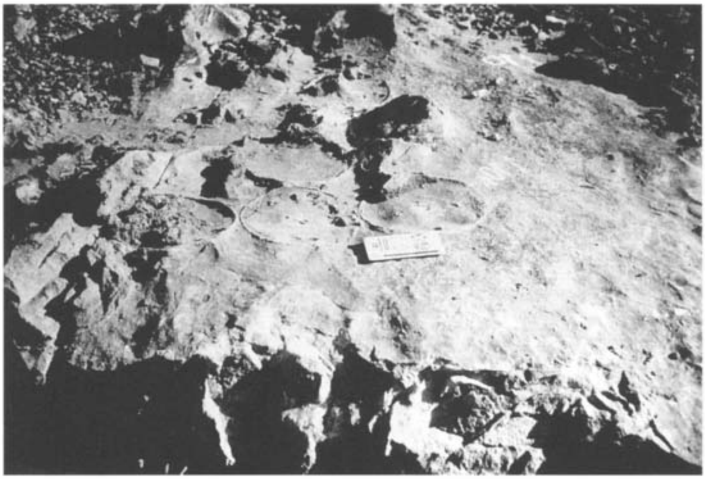

Figure 7a. Sauropod egg clutch with complete eggs. Locality: Lower Limestone, Balasinor, Lower Narmada region.

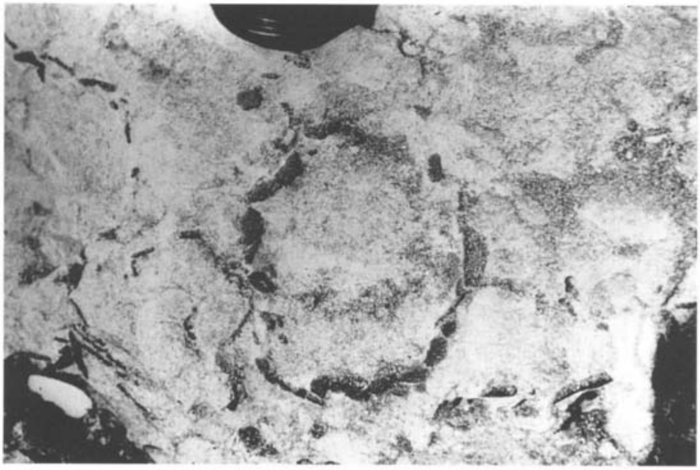

Figure 7b. Sauropod egg clutch with fragmented eggs. Outline of the eggs is discernible despite brecciation. Locality: Lower Limestone, Lametaghat.

3.2. Nagpur region

In the Chandrapur and Nagpur districts, detailed descriptions of the Lameta Formation have been provided from the Nand-Dongargaon Basin by Mohabey et al. [12] and Mohabey [13]. The Lameta Formation is exposed for a distance of 40 km between the Nand area in the north and the Dongargaon-Pisdura area in the south (Figure 1).

To enable comparisons with those of the Jabalpur region, a summary of the data given in Mohabey [13] is provided below. These beds are 20 m thick and are interpreted to have formed in alluvial-limnic environments under semi-arid conditions with seasonal climatic changes. Lithological characters of the Lameta facies deposited in different areas are quite variable. In the Nand-Dongargaon Basin, Mohabey [13] identified the following lithofacies with characteristic fossil assemblages: a) overbank red and green silty clays, b) channel sandstone, c) lacustrine cream and yellow laminated clays intercalated with limestone and marlite beds, and d) paludal grey marls.

The overbank facies are represented by red and green, massive to rarely laminated clays up to 6 m thick with lenses of channel sandstone. Clays mostly comprise montmorillonite, palygorskite, quartz, and calcite. Skeletal remains of sauropods associated with coprolites, chelonia, *Physa* and *Unio deccanensis* occur occasionally as lag deposits [13]. Other fossils found in these clays include reworked teeth and scales of fishes, *Lepodites*, *Lepisosteus*, *Pycnodus*, *Dasyatis*, *Igdabatis*, pharyngeal teeth (*Pycnodus*, *Stephanodus libicus*); Charophyta (*Platychara*, *Microchara*, *Peckichara varians*, *Stephanochara*, *Nemegtichara*); Ostracoda (*Paracypretta*, *Cytheridella*, *Cyclocypris*, *Cypria*), gastropoda; (*Physa*, *Melanea*, *Paludina*, *Valvata*, *Lymnaea*, *Turritella*), bivalves (*Unio* and *Corbicula*).

The channel sandstones consist of 2-6 m thick sandstones with occasional trough cross-bedding. The sandstone facies underlies, overlies and intercalates with the overbank facies [13]. Titanosaurid nest sites [27] occur in the calcretised sandstones. They also contain reworked dinosaur bones, teeth and scales of fishes, charophytes, ostracods and gastropods.

Thin grey marls within the overbank red clays represent paludal (backswamps) conditions. Desiccation features are common. Blooms of charophytes are associated with clusters of *Araucarites* seeds, impressions of palm and dicot leaf, ostracods and gastropods. The most predominantly occurring lacustrine facies consists of alternations of yellow and earthy brown shales intercalated with marlites and limestones which may be up to 8 m thick [12, 13]. Intercalated calcareous lithofacies exhibit wave-rippled structure. Thin layers of septarian concretions are locally present. This facies has a characteristic fish-ostracod-gastropod assemblage.

In addition to these major lithofacies, Mohabey [13] also recognised a very persistent palaeosol horizon in the red overbank clays and calcrete profiles in the upper parts of channel sandstones.

Dominance of overbank red clays has been used to interpret a well drained alluvial-plain with a major southwesterly flowing meandering river [13]. Intertonguing relationship of overbank red clays and lacustrine clays suggest that flood plain deposits overlapped the dried up lake surfaces. In places, the lakes developed playa conditions during extended dry seasons [13].

4. SEQUENCES AND PALAEOENVIRONMENTS OF INTERTRAPPEAN BEDS

The Intertrappean Beds are commonly made up of fossiliferous argillaceous and calcareous lithofacies which are mainly the deposits of small inland lakes. Also, based on foraminiferal evidence, some of the exposures in the vicinity of the East Coast have been interpreted to be outer shelf to inner shelf/upper slope marine deposits [28, 29].

In recent years two sections of the Intertrappean sequences have attracted particular notice. These are the Anjar Intertrappean Beds because of the report [30] of an iridium anomaly from them, and the Intertrappean Beds of Naskal, Andhra Pradesh because of the recovery of Late Cretaceous mammals [31, 32]. At Naskal, Prasad and Khajuria [32] suggested on the basis of sedimentological facies, habitats of flora and fauna, and accumulation patterns of microvertebrate assemblages, that the Intertrappean sequences were deposited in shallow alkaline lakes that were intermittently subjected to subaerial exposure. The predominant occurrence of freshwater lacustrine elements in the fauna and flora suggests deposition in shallow floodplain alkaline lakes.

The sequence at Naskal consists of gleyed and mottled mudstone, yellow mudstone, marl, chert and calcareous mudstone. The marls contain, in addition to vertebrate faunas, charophytes, gastropods, and ostracods. Subaerial exposure of floodplain-lacustrine sediments resulted in two phases of palaeosol development - the first phase being represented by gleysols and inceptisols, and the second phase by entisols and calcisols. Interestingly, this solitary mammal-bearing Intertrappean locality in India represents an autochthonous assemblage. Pronounced drought conditions possibly led to concentration of terrestrial animals including mammals near the drying lakes [32].

Like the Intertrappean sequence of Naskal, the Anjar Intertrappean Beds occupy a special significance, and have been the subject of recent studies [14, 30, 33, 34]. After Ghevariya [33] and Bajpai et al. [34, 35] recognised their importance for dinosaur eggshells, Bhandari et al. [30] reported three limonitic layers with iridium concentrations of 1200 pg/g well above that found in associated sediments and basalts (80 to 200 pg/g). Venkatesan et al. [36] using the $^{40}Ar/^{39}Ar$ method gave plateau ages of 65.2 ± 0.6 Ma and 64.9 ± 0.8 Ma for the lower flows, and 65.7 ± 0.7 Ma for the upper basalts. Importantly, Bajpai [37] reported the presence of dinosaur bones, microvertebrates, ostracods, molluscs, and spores. Microvertebrates include ornithoid (? Theropod) eggshells, fish teeth (*Igdabatis*) and otoliths/Serranidae and Aridae;

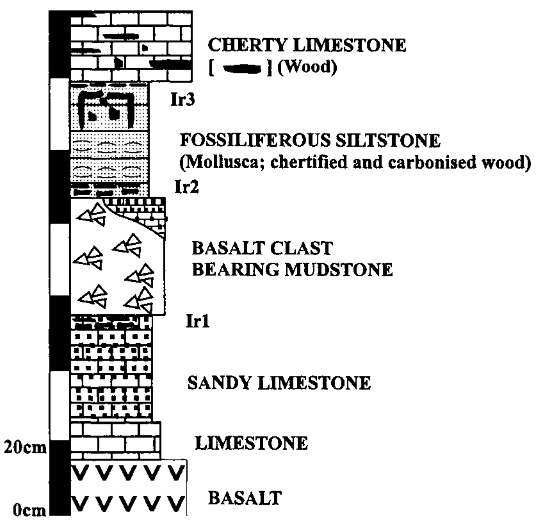

Figure 8. Litholog showing vertical facies association of a part of the Intertrappean interval at Anjar.
Note the three Ir enriched layers.

ostracods include *Altanicypris* and *Mongolianella*; plant microfossils include the Maastrichtian spore *Gabonisporites*, [37]. The Iridium layers at Anjar do not correspond to palaeontological extinction; their chemo-stratigraphic implications are therefore not properly understood despite their proposed connection to the K/T boundary [30].

The sequence at Anjar consists of cherty limestone and shales (Figure 8). Shale intervals are both laminated and massive. The shale beds have limonitic patches and stringers which have shown an iridium anomaly [30]. In places, the shales are highly fossiliferous and consist of molluscan shell debris. Mineralogically, the shales consist of Mg clay sepiolite with minor amount of smectite [14].

White to buff limestone beds show sharp contacts with the shale facies. Limestones are hard and, in places, show patches of black chert. They consist of peloids, microspar cement, mud intraclasts, bioclasts of microgastropods and molluscan shell hash [14].

The argillaceous and calcareous facies contain bioclasts, peloids, and faunal elements such as gastropods and bivalves which together suggest deposition in the littoral zone of alkaline lakes. Khadkikar et al. [14] have argued that the local climate during the formation of lake deposits was semi-arid, and possibly the result of mock aridity [38].

5. DISCUSSION

5.1. Regional palaeogeography

The depositional conditions outlined for the Bagh Beds, Lameta Beds, and Intertrappean sequences show that Late Cretaceous sedimentary environments in the central and western Indian region were variable and complex. The distribution of biotas and depositional setting of the Lameta and Bagh Beds in a regional context reveal that marine influences were confined only to the Lower Narmada Basin (Figure 1). The clastic facies overlying the Bagh carbonate lithofacies show definite estuarine characters. Although these clastic facies have been referred to the Lameta Formation [17], their stratigraphic affinity is uncertain. The regionally persistent dinosaur eggshell-bearing sandy carbonate lithofacies overlying the Bagh carbonate sequences in the Lower Narmada Basin is invariably referred to the Lameta Formation [17, 22, 39]. Most recent studies ascribe this to a freshwater lithofacies variably affected by pedogenesis [5, 19, 27].

Temporally, the Lower Narmada basin shows an estuarine facies (Nimar Sandstone) followed by a transgressive-regressive cycle (Nodular Limestone and Coralline Limestone) followed by an estuarine clastic facies (white/green sandstone), and eventually terminated by a sandy carbonate (calcrete) representing subaerial exposure conditions.

Facies distribution patterns reveal the absence of marine sequences in the easterly and southerly located region between Indore and Nagpur – a huge tract of peninsular India. In the Nagpur region, recent works of Mohabey et al. [12] and Mohabey [13] have stressed the presence of freshwater faunal elements as opposed to mixed faunal elements (marine and terrestrial) reported by earlier workers, particularly from the Intertrappean sequences [40, 41, 42]. Also a detailed work on biotas of the Anjar Intertrappean Beds [37] has stressed the presence of freshwater elements.

This review of Late Cretaceous environments suggests the following palaeogeographic scenario in Central India:

a) an epicontinental seaway advanced from the west during the Turonian (?) and inundated areas beyond Dhar but not to the east of Indore; marine influences persisted in the estu-

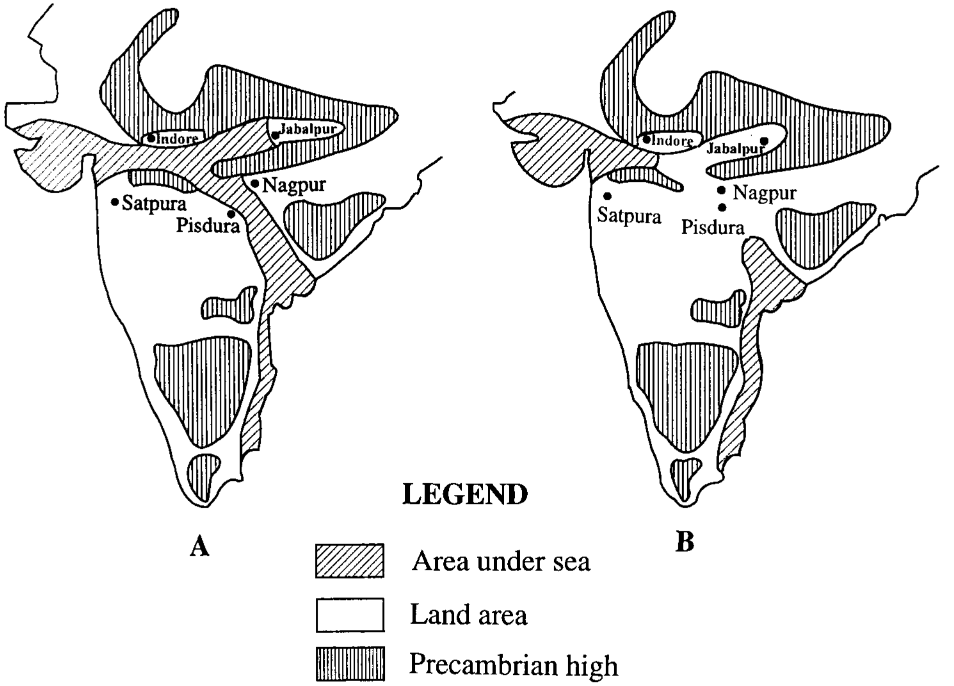

Figure 9. Maps to show the interpretations of distribution of land and sea during the Late Cretaceous. A: after Sahni [42] and B: after Sood [15].

b) arine green/white sandstones (of uncertain stratigraphic affinity) that overlie the Bagh carbonate lithofacies, widespread subaerial exposure conditions persisted in the entire region including the Lower Narmada Basin during the accumulation of the Maastrichtian Lameta Beds; fluvial /palustrine/sheetwash conditions predominated in the Jabalpur sub-region; palustrine conditions occurred throughout the region; and alluvial-limnic conditions were predominant in the Nagpur sub-region.

Sahni [42] proposed a single transgression during the Turonian along the Narmada-Son trough and the Godavari-Pranhita Trough along a seaway connecting the southeastern and western coast of India (Figure 9). Presence of alluvial/lacustrine/palustrine conditions in Nand and Dongargaon (Godavari-Pranhita Trough) and Jabalpur (Narmada-Son Trough), and the absence of confirmed marine facies in these regions negates the existence of a seaway connecting the south Indian Cretaceous areas with the western marine areas of the Narmada Trough (Figure 9).

5.2. Tectonic setting and evolving sedimentation patterns

Figure 10 shows a reconstruction of eastern Gondwana at 120 Ma with positions of the active rift zones [43]. Sediment dispersal in the Central Indian rift zone was active as evi-

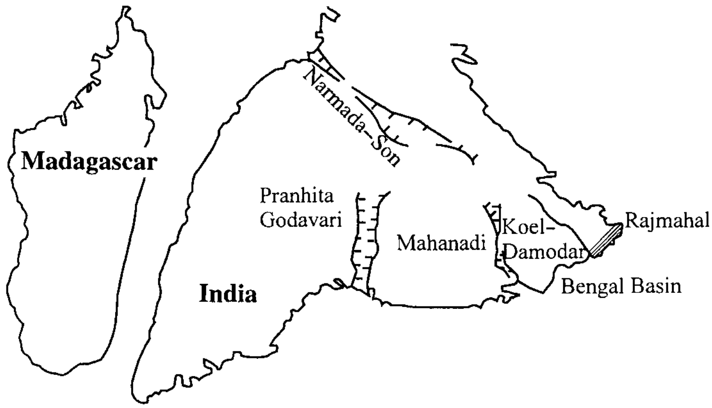

Figure 10. Reconstruction of eastern Gondwana at 120 Ma showing the positions of rift zones and the thermally perturbed Rajmahal region (after Kent [43]).

denced by the upper Gondwana sedimentation (Jabalpur Group). Formation/reactivation of discrete rift structure in the region was possibly related with the initiation of a thermal anomaly in the asthenosphere [43], as indicated by the outpouring of Rajmahal lava flows in the Early Cretaceous [44]. Early Cretaceous strata (Nimar Group) show marine influences in the west whereas only continental facies are found in the inland areas (Jabalpur Formation). With further northward movement, the rifted central and western parts of the Indian lithosphere came under the influence of another thermally perturbed region, the Reunion.

The western region of the Narmada Trough was also affected by plume-related surface uplift and resulted in the withdrawal of the sea. Later, coeval with early Deccan volcanism, the Lameta sediments formed on available surfaces in the rift zone. These included the rift-related Gondwana and Bagh Beds as well as older Precambrian sequences. When a plume rises beneath a continent, the dynamic and thermal effects produce both phases of uplift and fracturing of domal uplifts [45]. Volcanism occurs in these terrains that are subject to uplift and faulting, the Deccan being an example of such a geodynamic terrain [46]. In such settings, inland continental areas minimally effected by faulting would be most suitable for the occurrence of widespread subaerial exposure conditions. Relief/topographic differentiation should be minimal in such areas and lacustrine/sheetwash deposition would be significant. Contrarily, those areas where uplift related gravitational forces are high and related fault subsidence is significant would have maximum relief and topographic differentiation. Such areas would be favourable for the accumulation of alluvial-lacustrine facies. In Central India, alluvial-lacustrine facies are predominant in the Nagpur sub-region whereas palustrine/sheetwash deposits are predominant in the Jabalpur-Indore sector (Figure 1). This reflects more topographic differentiation in the former areas as opposed to the latter. Although such topographic differentiation may reflect regional differences in the erosional relief, it may also be the result of gravitationally induced fault-related subsidence.

6. CONCLUSIONS

Analysis of patterns of environmental change reveals a trangressive carbonate lithofacies (Turonian?) in the Lower Narmada Basin which is generally followed by dinosaur eggshell-bearing calcareous sandstone/sandy brecciated calcrete (Maastrichtian) representing wide-spread subaerial exposure environments. This unconformable contact marks the westward withdrawal of sea fom the inland embayment; possibly influenced by plume-related surface uplift that preceded the Deccan flood basalt volcanism.

In most inland parts of the intracratonic rifts of the Indian lithosphere, Early Cretaceous Gondwana fluvial facies were unconformably overlain by the Maastrichtian Infratrappean Beds that consist predominantly of palustrine and sheetwash deposits, variably effected by pedogenesis. This dinosaur eggshell-bearing palaeoregolith was broadly coeval with the basal Deccan lava flows.

In the Nagpur region, the Infra-trappean Beds are predominantly made up of alluvial-lacustrine facies. Such areas of the Cretaceous Indian lithosphere were possibly effected by surface uplift and related fracturing/faulting; the latter processes produced a terrain with sufficient relief and topographic differentiation to support alluvial-shallow alkaline lake systems.

Continental facies, coeval with the basal lavas, in places, show features suggestive of semi-aridity; climatic semi-aridity may have been enhanced by episodic volcanic activity.

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Scientific achievements of IGCP-350 "Environmental and Biological Change in East and South Asia during the Cretaceous": an overview

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The major scientific achievements of the activity of IGCP 350 (1993-1998) have been summarized as regards (1) stratigraphy and correlation, (2) plume-related events, (3) biological aspects of environmental change, (4) physical aspects of environmental change, (5) chemical aspects of environmental change, (6) climate enviroments, and (7) natural resources, all of which were the main topics of researches of this project.

1. INTRODUCTION

IGCP project 350 aimed to better understand environmental and biological changes in eastern and southern Asia (including Southeast Asia) during the Cretaceous Period by co-ordinated inter- and multidisciplinary research in stratigraphy, sedimentology, paleontology, petrology, geochemistry, tectonics and geophysics [1]. In order to confirm synchroneity of environmental change events, it has been critical to assess the combined data from bio-, litho-, magneto- and chemo-stratigraphic sequences to provide a working standard stratigraphy. Biological change has been compared with this standard.

It is generally believed that unique Cretaceous environments are well documented in rocks exposed widely both on land and under the sea in eastern and southern parts of Asia (Figure 1). Furthermore, in a broader geographic context, geological interactions between the proto-Pacific and the Tethyan-Boreal regions during the Cretaceous are most important [2]. It has been anticipated, therefore, that high quality data on environments specific to these regions will be forthcoming. In particular, Cretaceous plume tectonic regimes played an important role in the superplume volcanic events in the proto-Pacific that are a key in understanding the

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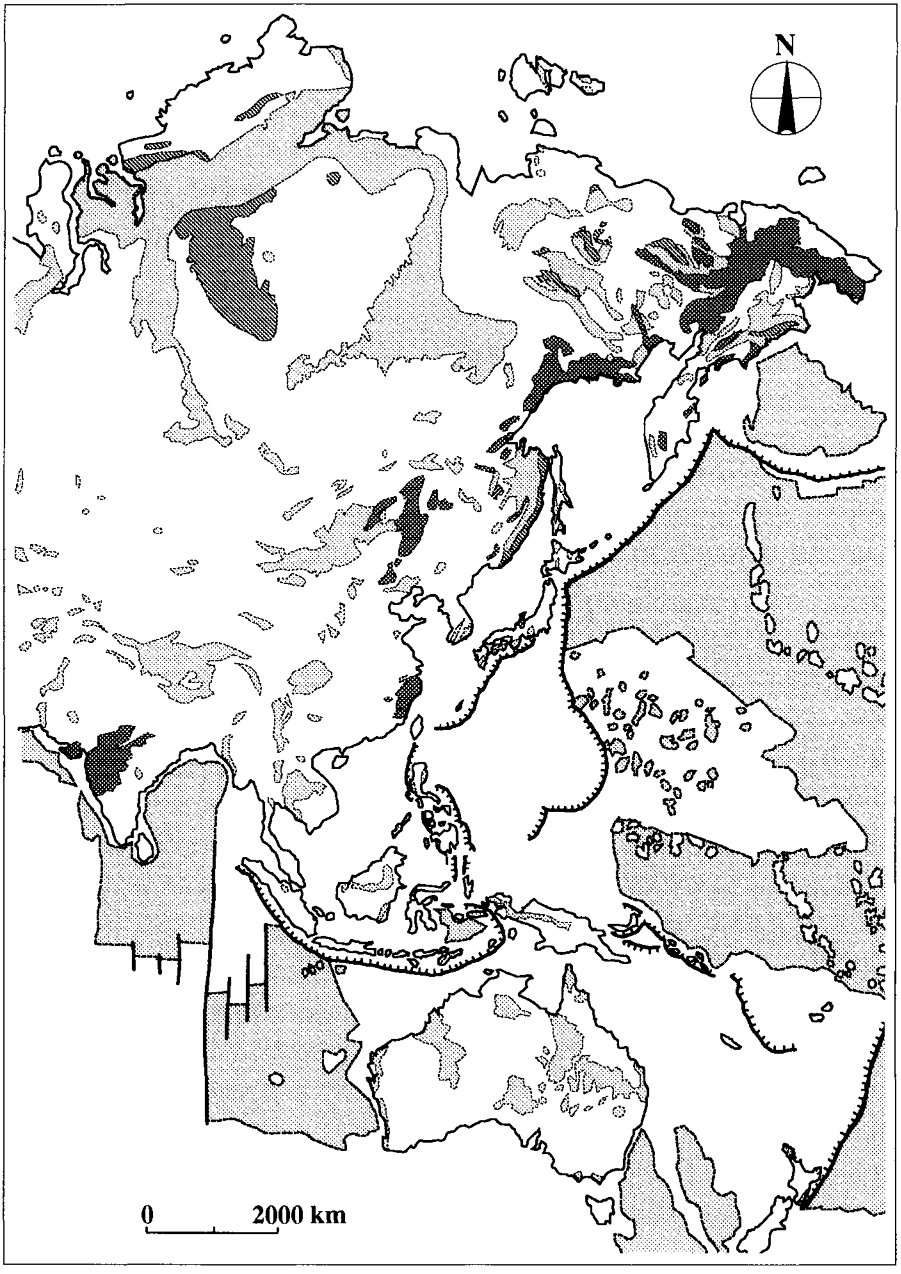

Figure 1. Map of Asia, Oceania and the West Pacific, showing the distribution of the Cretaceous System (shaded) (compiled from CGMW/UNESCO [4]). Undivided Mesozoic is also included on land. Dark shaded is volcanic formations. Scale is for the Equator region.

Cretaceous environments in East Asia [3].

Integrated results from research in the regions covered by this project have added plenty of important information on Cretaceous environmental features. This paper summarizes the major scientific achievements accomplished under the auspices of IGCP 350 from 1993 to 1998.

2. MAJOR SCIENTIFIC RESULTS

2.1. Stratigraphy and correlation

Stratigraphic correlation is critical to identifying the synchroneity of various geological and biological events. Much effort has been expended on this activity in China, India, Indonesia, Japan, Korea, Malaysia, Mongolia, New Zealand, Pakistan, Philippines, Russia, Spain, Thailand and Vietnam. The following are the representatives of those results:

2.1.1. Establishment of standard bio-stratigraphy, magneto-stratigraphy and chemo-stratigraphy

The most important work on the establishment and integration of mega-, micro- and magneto-stratigraphy of the Upper Cretaceous has been carried out in Japan [5, 6]. Foraminifer biostratigraphy in India [7], Philippines [8] and China [9], radiolarian biostratigraphy in Japan [10, 11, 12] and Russia [13], conchostracan biostratigraphy in China [14], ostracod biostratigraphy in India [15], phyto-biostratigraphy in Russia [16], and chemostratigraphy in Japan [17, 18] are also important.

2.2.2. Correlation between marine and nonmarine stratigraphy

Some attempts have been made to correlate marine and nonmarine Lower Cretaceous sequences in China [19, 20], Japan [21] and Russia [22] on the basis of the molluscan, radiolarian, conchostracan and ostracod biostratigraphic schemes. In particular, conchostracan zonation established in China has successfully been applied to the Japanese nonmarine Cretaceous [23].

2.2.3. Correlation between the proto-Pacific, Tethyan and Boreal realms

Great effort has been rendered in China [9], Japan [5, 21], India [24], New Zealand [25], Pakistan [26], Russia [27] and Spain [28, 29] by using molluscan, nannofossil, foraminiferal and radiolarian biostratigraphy.

2.2.4. J-K boundary

The marine successions including the J/K boundary horizon have been examined in the Japanese Islands and in the western Pacific on the basis of a newly established radiolarian zonal scheme [10, 30], which has been approved to be useful in variable facies sequences. In the Amuro-Zeyskaya and Bureinskaya basins in southeastern Russia, somewhat stable environments were suggested across the boundary [31]. In the marine sequence in South Primorye, near Vladivostok, a mixed fauna of the Boreal and Tethyan realms was found [27, 32]. This is of great importance to solve problems with paleobiogeography and migration of faunas between the Boreal and Tethyan regions during the Late Jurassic to Early Cretaceous time.

2.2.5. Cenomanian-Turonian boundary

Chemical features and the chronology of the C/T boundary were studied intensively in the Yezo Supergroup in Hokkaido, Japan [17, 18, 33]. Calcareous nannofossil study at the C/T boundary at Ganuza in the Navarra province in northern Spain [28, 29] has revealed the significant decrease in oceanic primary productivity related to the C/T boundary event. Another significant contribution is the establishment of the reference section for the C/T boundary in Europe [34].

2.2.6. Santonian-Campanian boundary

An accurate correlation of the Santonian-Campanian transition in the Yezo Supergroup in Hokkaido, Japan, has been established with combined bio- and magneto-stratigraphy [6], which revealed that the boundary is situated in the marine magnetic normal polarity chron 34.

2.2.7. K-T boundary

The K-T boundary in the nonmarine sequence has been carefully checked in the Jiangsu (N China), Guangdong (SE China), Jianghan (southern Interior China) and Wuyun (NE China) areas [19]. In India, the K-T boundary was intensively studied in the Deccan Inter- trappean sequence, where not only the discovery of an iridium anomaly [35] but also new data of palynology, ostracod and vertebrate assemblages [36] were obtained.

2.3. Plume-related events

The huge magmatic development in East Asia in the Early Cretaceous cannot be explained solely by the magmatism related to plate subduction, because they are not wholly correlated with subduction phases [37]. Notwithstanding the development of active magmatism in the Early Cretaceous, no subduction took place at least along the segment of Southwest Japan in this period [11, 12, 38]. This discrepancy has been explained by plume-related magmatism [37] based on the Yano and Wu's model [39, 40] (Figure 2).

2.4. Biological aspects of environmental change

2.4.1. Diversity of dinosaurs and taphonomic study of their habitats

Many interesting studies have been carried out; in particular, on the diversity of dinosaurs in Thailand and their origin [41], migration of dinosaurs in Mongolia [42], taphonomy of dinosaur eggshells in India [43, 44]. It is noteworthy that the occurrence of very thin eggshells of dinosaurs from the Deccan sequence, India, suggests extreme environmental stress due to Deccan volcanic activity [44]. The social behavior of dinosaurs has also been studied, mainly based on the analysis of trackways [41, 45, 46, 47, 48, 49].

2.4.2. Vegetation, phytogeography and climatological environments

Diversity of Cretaceous vegetation has been presented with respect to China [50, 51], India [36], Japan [52, 53, 54] and Russia [16]. The paleophytogeography in the Early Cretaceous of Japan reveals a Ryoseki-type flora that grew under tropical or subtropical conditions with a yearly arid season is characteristic of the Outer Zone of Southwest Japan and Southeast China; the Tetori-type flora grown under the warm-temperature belt is of the Inner Zone of Southwest Japan, Korean Peninsula, Northeast China and most of Siberia; and the Mixed-type flora is found in the southern tip of the Korean Peninsula, Yellow River area in China and Southern Primorye in Sikhote-Alin [53, 54].

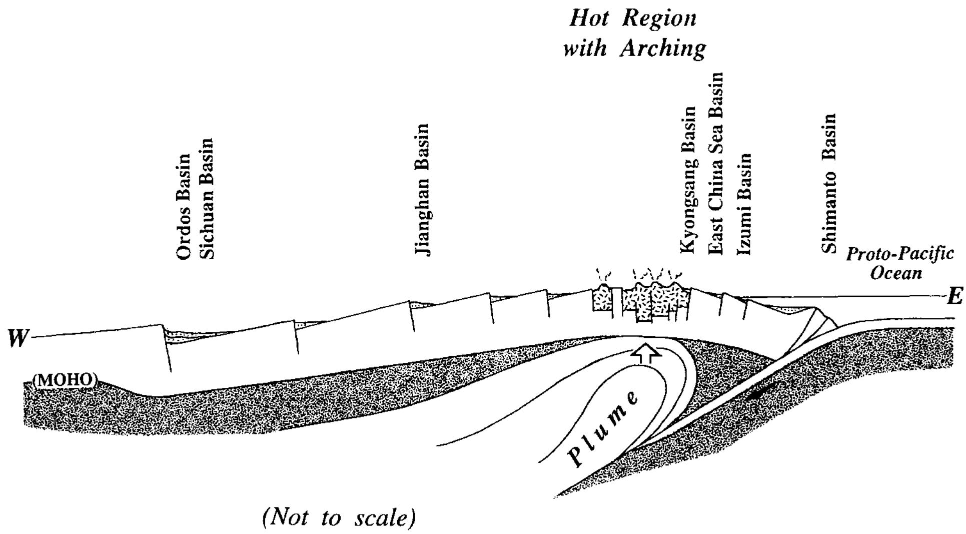

Figure 2. Yano and Wu's model showing the development of rift basins closely related to plume activity at the East Asian continental margin (adapted from Okada [37]).

2.4.3. Bio-diversity analysis

Much significant work has been carried out on conchostracans [14], molluscs [55, 56, 57, 58], fish [59], dinosaurs [41, 43, 46, 47, 48, 49], insects [60, 61], and plants [16, 50, 51, 52, 53, 54]. In particular, the largest fossil insect fauna in the world has been added in excellent preservation from the Early Cretaceous Laiyang basin, Shandong Province, NE China, which consists of 15 orders, over 100 families, nearly 400 genera, and 750 species [61].

2.4.4. Palaeogeography based on faunal data

Japanese and Russian scientists have attempted paleogeographic reconstructions mainly for the Early Cretaceous in East Asia [21, 32, 62]. In particular, Matsukawa et al. [21] presented oceanographic current systems around the Japan and Sikhote-Alin areas during the Early Barremian, showing the southward currents from the high latitudes near the coast and the northward currents from the Equatorial regions off the coast.

2.4.5. Origin of primitive birds

A very important discovery of the excellently preserved feathered dinosaurs *Sinosauropteryx prima* from the uppermost Jurassic or lowermost Cretaceous Yixian Formation of the Jehol Group in western Liaoning, NE China [63, 64]. These creatures have contributed much to our understanding of the early evolution of birds [63]. The exact age determination of the strata containing these fossils remains under study (oral communication, Chen Pei-ji).

2.4.6. Origin of angiosperms

The site that contains some of the earliest of angiosperms [50] is located near the locality of the above-mentioned primitive birds [65]. The present angiosperm megafossils come from the Berriasian-Hauterivian (or early Barremian) Chengzihe Formation at Jixi, eastern Hei- longjiang Province [66], NE China, which is younger than the horizon of the above-mentioned primitive birds. Sun and Dilcher [50] advocate that "the East Asia region, including northeastern China and Mongolia, would have been the original center of the angiosperms in the world, from where the earliest angiosperms had evolved and radiated to the Eastern Gondwanaland and other regions".

2.5. Physical aspects of environmental change

2.5.1. Genetic environments of red beds

Cretaceous red beds are widely distributed in Asia. A useful review of the origin of red beds in East Asia has been provided by Japanese workers [67, 68]. Red beds in East Asia are divided into two groups: those with and those without evaporites [67, 68]. The evaporitic red beds occupy the basins succeeded from Jurassic salt-lakes, and the non-evaporitic ones are distributed in South Korea and Japan.

Miki and Nakamuta [67] discussed the genetic environments of red beds in two ways: (1) origin of iron and (2) coloration process. As to the origin of iron, they clarified an intimate relationship between iron in the red beds and iron-bearing minerals in the basement rocks. Thus, iron has been originated by weathering of iron-bearing silicate and iron-oxide minerals in the source rocks, with subsequent transportation of material together with clay fractions to the depositional sites. They were later transformed to the pigmenting hematite.

Weathering and decomposition of source rocks to release and oxidize iron occurred under moderate and sub-humid conditions. Iron substances transported from source areas to the freshwater depositional basins were oxidized in high Eh conditions. During red-bed sedimentation, accompanied plant materials caused a local reducing environment and partial bleaching of color.

They further stated that hot and dry climates under which evaporitic minerals formed in contemporaneous formations in southern China and Indochina may have played a less important role in red-bed formation in Central Kyushu, Japan.

2.5.2. Detrital chromian spinels and their tectonic implication

Hisada et al. [69, 70] have proposed detrital chromian spinels as a useful tool for detecting mafic-ultramafic rock provenance, which were incorporated into suture zones during collisions and/or accretionary processes. In this way, timing of suturing of Cretaceous basins in many orogenic belts in East Asia has been clarified [69, 70, 71, 72, 73].

2.5.3. Tectonic environments of basin development, especially in relation to strike-slip fault movements

Relationship between large scale characteristic tectonic features of East Asia, such as the Tan-Lu Fault system, and basin development have been discussed by Chinese and Japanese scientists in terms of oblique plate subduction (Figure 3) [38, 74, 75, 76, 77]. The strike-slip basins in Southwest Japan indicate eastward-younging sedimentation [37, 75], which is closely correlated with the magmatism of eastward migration [37] due to oblique plate subduction.

In addition, in the Tethyan region, tectonic environments of basin development have been

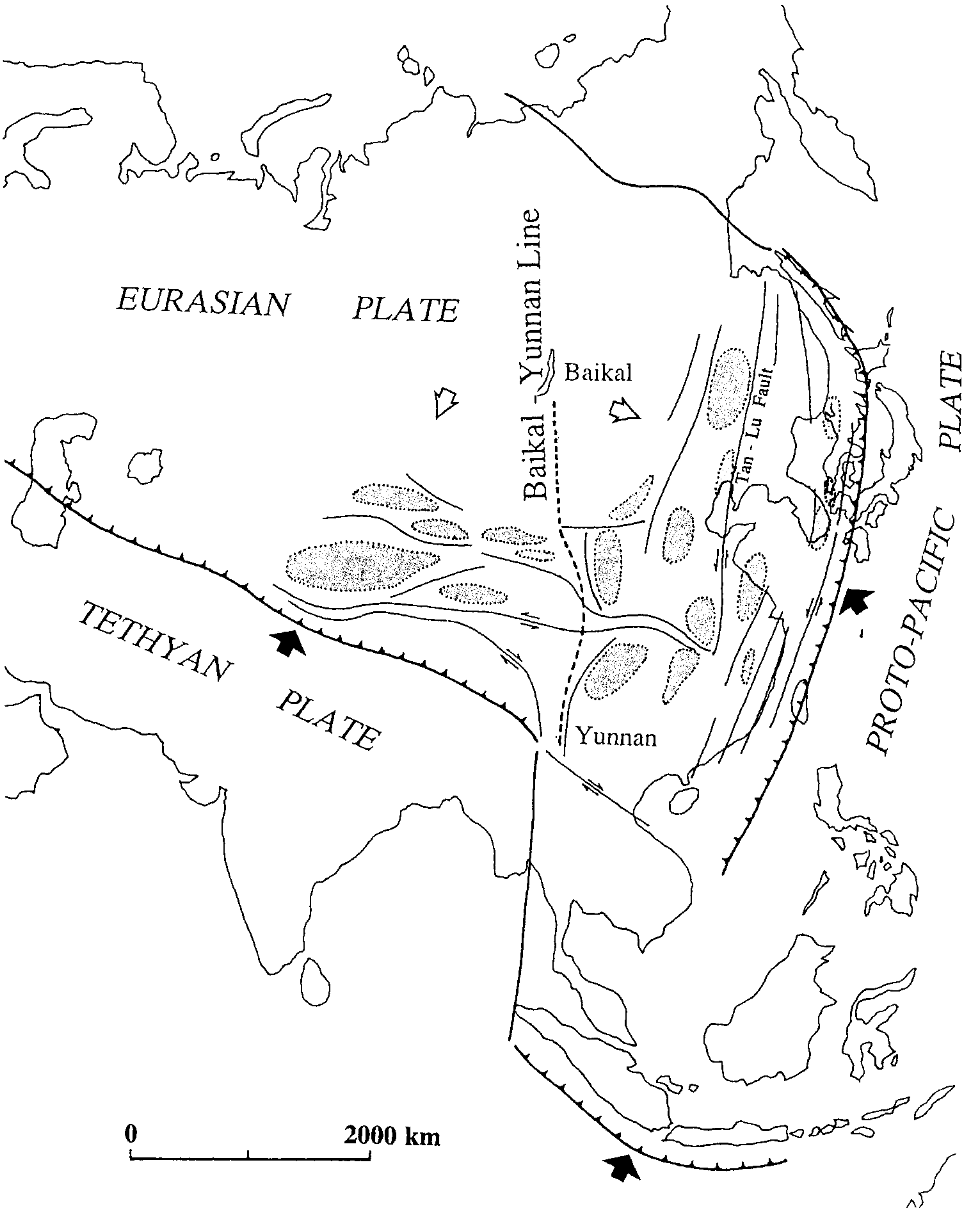

Figure 3. Tectonic framework and basin arrangement on the Cretaceous Eurasian Plate (adapted from Okada [37]). Solid arrows: direction of plate movement, open arrows: direction of regional stress. Scale is for the Equator region.

discussed in Pakistan [26] and India [78], where sequential rifting events produced sedimentary basins on the supercontinent Pangea, and in Sarawak of Malaysia [79] a plate tectonic evolution of the Sarawak and Kalimantan areas was summarized, showing a collision event of the West Sarawak block with the West Borneo basement and a development of a large Cretaceous basin behind the collision suture.

2.5.4. Sequence stratigraphy and sea-level change

Interesting data of sequence stratigraphy have been identified in some Japanese Cretaceous sequences [80, 81, 82, 83]. It is noteworthy that extraordinarily high rates of clastic sedimentation as high as 500 m/m.y. to 4000 m/m.y. were encountered in the Japanese Islands during the peak stage of the global transgression in Cenomanian and Campanian times [82]. This implies that regional tectonic activity is more important in clastic sedimentation in orogenic belts than the first- to third-order global sea-level changes. Such high rates of clastic sedimentation are expected irrespective of sea-level changes, if high topographic relieves and thier high rates of weathering and erosion under wet climate are available.

Another important discovery is Barremian ammonite-bearing marine strata in Heilongjiang Province, NE China [21], which implies the regional marine transgression in the Barremian time.

2.5.5. Provenance and depositional environments

Many case studies of provenance analysis of Cretaceous sediments have been presented in India [84], Japan [73, 85, 86, 87], Korea [88, 89], and Russia [90]. Further attempts have been made to reconstruct paleoenvironments on basinal to regional scales, based on sedimentological and biological data in India [91, 92] and Japan [46, 93].

2.5.6. Magmatism

The migration model of magmatism has been developed to the Mesozoic igneous rocks of the East Eurasia margin [94, 95, 96]. In addition, some new data on the nature of magmatism in Far East Russia [97], North Korea [98] and the Philippines [99] have been added.

2.5.7. Ridge subduction

Kiminami et al. [100] and Kinoshita [94, 95, 96] have shown the relation between the oblique subduction of ridges and the migration of magmatism. This notion was advocated to have been effective for migration of sedimentary basins [75]. It is probable, however, that the magmatism related to ridge subduction may have been true only in a restricted segment of the East Asia continental margin [37], for example, in the Southwest Japan arc.

The thermal effect on the Shimanto accretionary prism due to the ridge subduction was analyzed by Sakaguchi [101].

2.6. Chemical aspects of environmental change

2.6.1. Isotopic composition and environments

Indian, Japanese and Korean workers have presented important isotopic data in relation to chemical environments [17, 102, 103, 104]. In particular, carbon and oxygen isotope composition in Early Cretaceous paleosols in Korea gives a new data on the CO^2 composition of the atmosphere at that time [105].

2.6.2. Oceanic anoxic events (OAE 1 and OAE 2)

Lithological and chemical features of OAE 1 (Aptian/Albian Anoxic Event) and OAE 2 (Cenomanian/Turonian Anoxic Event) have been clarified in the Yezo Supergroup in Hokkaido, Japan [17, 106, 107, 108], clarifying that the depositional site was under reducing conditions that prevailed from the outer shelf to the upper shelf slope, and that calcareous organisms flourished before and after the oceanic anoxic event.

2.6.3. Chemical features

Some chemical data of major and trace elements have been presented on Japanese Cretaceous sediments [109, 110]. One example comes from the Lower Cretaceous nonmarine sediments in Southwest Japan [109], delineating paleo-lake environments, and the other from the Upper Cretaceous pelagic sediments in the Shimanto Belt [110].

2.7. Climate environments

2.7.1. Sedimentological aspect

Sedimentological and climatic aspects were discussed on Japanese and Korean red beds [67, 68, 111], suggesting semiarid to semihumid climate or alternating hot/dry and/or hot/humid climates [68].

2.7.2. Biological aspect

Climatic and paleogeographic environments through faunal analysis have been discussed in China [19], India [84, 91, 92], Japan [21, 53, 54], Mongolia [42, 46, 112, 113] and Russia [114]. A good summary of climatic environments of Asia based on vegetation characteristics has been given by Ohana and Kimura [53], Herman and Spicer [114] and Spicer et al. [115] . In particular, Spicer et al. [115] depict the Cretaceous phytogeography and climate of Asia with qualitative and quantitative data, and provide numerical climate model for the Cenomanian.

2.8. Natural resources

Coal, oil, and varieties of metallic ores are in huge abundance in the Asian Cretaceous, among which potentiality of petroleum and gas resources has been discussed by Pakistan workers [26]. Depositional systems of uranium deposits in India was deciphered by D'Cruz et al. [116].

3. CONCLUDING REMARKS

As summarized above, the project IGCP 350 has undertaken significant research along defined but broad guidelines. The most important are the establishment of correlation of many geological events by means of integrated bio-, litho-, magneto- and chemo-stratigraphy, and the recognition of the cessation of accretion at least along the segment of the Southwest Japan arc in the Early Cretaceous. The latter fact suggests that the intense igneous activity in the Early Cretaceous may be ascribed to plume activity.

Environmental features of the regions covered by IGCP 350 have been assessed in many respects, as given separately in the preceding sections in this volume. It is hoped that the regional study in Southeast Asia and Russian Far East will develop further in the near future, since these regions are the key areas between the Tethys and the proto-Pacific as well as between the proto-Pacific and the Boreal [3].

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Table 1

Stratigraphic correlation of Cretaceous sediments between regions

Time Scale: Age	Region. Subdiv.: Formation	(A) Altai: Great Lakes (I): Khosbayar [2]	(A) Altai: Valley of Lakes (II): Shuvalov [3]	(A) Altai: Valley of Lakes (II): Sinitsa [5]	(B) Khangai - Khentei: Arkhangai (III): Khosbayar [6]	(B) Khangai - Khentei: Onon (IV): Khosbayar [6]	(B) Khangai - Khentei: Choir - Nyalga (V): Khosbayar [6]	(B) Khangai - Khentei: Choir - Nyalga (V): Matsukawa et al. [7]	(B) Khangai - Khentei: Choibalsan (VI): Khosbayar [6]	(C) Gobi: Transaltai Gobi (VII): Shuvalov [3]	(C) Gobi: Umnogobi (Southern Gobi) (VIII): Shuvalov [3]	(C) Gobi: Umnogobi (Southern Gobi) (VIII): Lopatin [17]	(C) Gobi: Umnogobi (Southern Gobi) (VIII): Khand & Badamgarav [18]	(C) Gobi: Domogobi (Eastern Gobi) (IX): Martinson et al. [20]	(C) Gobi: Domogobi (Eastern Gobi) (IX): Shuvalov [3]	(C) Gobi: Tamtsag (X): Khosbayar [6]
Up. Paleoc.	Khashaat									Naranbulag						
Maastricht.	Nemegt									Nemegt	Nemegt			Javkhlant	Nemegt	
Campanian	Baruungoyot									Baruungoyot	Djadokhta		Bayanzag		Baruungoyot	
Coniacian-Santonian	Djadokhta															
Cenomanian Turonian	Bayanshiree									Bayanshiree	Bayanshiree		Bayanshiree	Bayanshiree	Bayanshiree	Bayanshiree
Albian Cenoman.	Baruunbayan		Baruunbayan										Baruunbayan	Sainshand	Baruunbayan	Sainshand
Aptian-Albian	Khukhteeg	Zereg	Khulsangol	Khulsangol; Boontsagaan Series: Kholboot Beds	Shankh	Bayankhaan	Uvdugkhudag; Choir		Aduunchuluu	Baruunbayan; Dushuul	Baruunbayan; Ulaandel	Sainshand; Khukhteeg	Khukhteeg	Khukhteeg	Khukhteeg	Zuunbayan
Hauterivian Barremian	Shinekhudag		Andaikhudag	Boontsagaan Series: Khurilt Beds; Shartolgoi Beds	Battsengel	Zaanshiree	Nyalga	Shinekhudag	Gurvanzagal			Shinekhudag; Manlai	Shinekhudag; Manlai	Shinekhudag	Shinekhudag	
Valanginian-Berriasian	Tsagaantsav	Gurvanereen	Undurukhaa	Undurkhin	Arkhangai	Khongork hairkhan	Kherlen	Tsagaantsav	Sumiinnuur; Bayantumen; Ochirkhureet		Tsagaangol	Mogoit	Tsagaantsav	Tsagaantsav	Tsagaantsav	Tsagaantsav
Tithonian-Kimmeridgian	Shariliin	Ikhesnuur	Toromkhon	Dundargalant Series: Shandrol Beds; Andakhudag Beds	Jargalant	Seruungol	Dorgot	Sharliin					Shariliin	Shariliin	Shariliin	

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Table 2

Stratigraphic position of representative **Cretaceous fossil assemblages**

System	Stage	Formation	Charophytes	Spores & Pollen	Plants: Krassilov [29]	Plants: Sodov [30]	Plants: Makulbekov & Kurzanov [32]	Molluscs	Ostracods	Dinosaurs	Conchostracan	Turtles	Lizards	Mammals
Cretaceous	Maastricht.	Nemegt						***Buginella buginica Mesolanistes***	***Talicypridea reticulata*** / ***Mongolocypris distributa***	*Tarbosaurus bataar Gallimimus bullatus Deinocheirus mirificus*	*Estherites Paleoleptestheria Liaestheria*	***Mongolochelus*** Lindholmemydidae Haichemydidae	***Slavoia darevskii Shinsauroides latipalatum***	*Buginbaatar transaltaiensis*
	Campanian	Baruungoyot	***Mongololiachara Mesochara***		*Nyssoidea mongolica*		*Ginkgo Platanus Trochodendroides Nelumbo Quercexia*	***Pseudohyria barungoiotica Buginella laevivalvata***	***Talicypridea abdarantica***	*Velociraptor mongoliensis Protoceratops andrewsi Mononychus olecranus Pinacosaurus grangeri*	?	***Adochida*** Nanhsiungchelyidae	***Priscagamu gobiensis Mongolocha mopsfacilis***	*Nemegtbaatar gobiensis Deltatheridium tardum Asiatherium reshetovi*
	Santonian, Coniacian	Djadokhta	?		*Bothrocaryum gobiense Nyssoidea mongolica*		?	***Plicatotrigonioides gobiensis***	***Gobiocypris tugrigensis***				***Mimeosaurus crassus Gobekkocretacicus Isodontosaurus gracilis***	*Kamptobaatar kuczynskii Zalambdalestes lechei Deltatheroides cretacicus Kennalestes gobiensis*
	Turonian, Cenoman.	Bayanshiree	***Atopochara multivolvis Caucasuella gulistanica***		?			***Pseudohyria cardiiformis***	***Lycopterocypris baishintsavica***	*Segnosaurus galbinensis Garudimimus brevipes*	*Sphaerograpta* sp. *Eulimnadia arjachlovae* / ?			?
	Albian	Baruunbayan	***Mesochara sainshandinica M. tuzsoni***		*Limnothetis Limnoniobe*	*Cyparacites anlajensis Potamogeton* sp.		***Campeloma yihsiensis Oxynaia sainshandica***	***Cypridea rostrata***	?	*Paleoleptestheria* sp.	?	?	
	Albian, Aptian	Khukhteeg	***Atopochara trivolvis Mesochara voluta M. tuzsoni***	***Foraminisporites assemmetricus Alisporites elongatus Abiespollenites* sp.**	*Baierella hastata Araucaria mongolica*	*Araucaria* sp. *Parataxodium jacutensis Ginkgo adiantoides Asplenium dicsonianum* / *Muscites ostracodiferas Dicksonia concinna Otozamites lacustris*		*Sainshandia ushlegica S. dushiulensis*	***Tsetsenia mira Trapeoidella khandae Cypridea acutituberculata Janinella tsaganensis***	*Harpymimus okladnikovi Iguanodon orientalis*	*Bairdestheria multa Estherites liushinensis*	***Hangaiemys leptis Asiachelys perforata*** / ***Hangaiemys hoburensis*** Macrobaenidae ***Moingolemys*** sp.		*Guchinodon khoburensis Eobaatarmagnus Gobiodon infinitus Arguimus khosbajari Kielantherium gobiensis Piakennalestes trofimovi*
	Barrem., Hauteriv.	Shinekhudag	***Aclistochara caii Raskyella* sp.**	***Cicatricosisporites australiensis Densoisporites velatus Pilosisporites trichopapillosus***	***Otozamites Pseudolarix erensis***	*Birisia onychioides Ruffordia geoppertii Otozamites* sp. *Araucarites* sp.			***Cypridea fasciculata***	?	*Bairdestheria memorabilis B. shinchudukensis*			
	Valang., Berrias.	Tsagaantsav			*Baiera manchurica*	*Cladophlebis Onychiopsis Nilssoniopteris denticulata*			***Cypridea unicostata***	*Psittacosauridae*	*Opsipolygrapta anferovi Bairdestheria enucleata*			

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