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Paleogeography and Sedimentation Settings during Permian–Triassic Reorganizations in Biosphere

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Abstract—Origination and growing aridity of Pangea and transition from glacial to ice-free climate led in the Permian–Early Triassic to substantial reorganization in all biosphere subsystems. Lithological–paleogeographic maps drawn for the late Sakmarian–early Artinskian and late Kazanian–early Toarcian periods of the Permian and for the Induan Age of the Early Triassic illustrate main paleogeographic changes that were related to the Pangea uplift, successive regression of epicontinental seas, disappearance of glacial sedimentation areas, progradation of arid and semiarid sedimentation, expansion and multiplication of closed basins, and reduced surficial but enhanced groundwater discharge into shelf seas surrounding Pangea. This distorted the global balance of mineral exchange between land and ocean and reduced the terrigenous influx of nutrients and primary productivity in ocean that could destruct the former feeding chains and represent one of possible factors responsible for mass extinction at the end of the Permian.

Key words: Permian, Triassic, paleogeography, sedimentation settings, Pangea, global mineral exchange, biosphere reorganizations.

1. INTRODUCTION

Permian and Early Triassic climatic changes and reorganizations in biosphere are most remarkable in the Phanerozoic history. They are manifested in transition from the glacial to ice-free climate, in growing aridity of Pangea (Parrish, 1993, 1995), and in mass extinction of marine and terrestrial faunas at the end of the Permian Period (Maxwell, 1989; Raup and Sepkoski, 1986; Erwin, 1995; Alekseev, 1998; Ochev and Shishkin, 1998; Ponomarenko and Sukacheva, 1998; and others). Substantial changes affected all subsystems of biosphere and were accompanied by global changes in Sr, C, and S isotopic ratios (Burke *et al.*, 1982; Baud *et al.*, 1989; Denison and Koepnik, 1995; Erwin, 1995a; Scholle, 1995).

The most intense biosphere reorganizations developed since the second half of the Early Permian until the Induan Age of the Early Triassic. They were preceded by durable paleogeographic and climatic changes, which should significantly influence sedimentation environments and, consequently, habitat conditions of organisms in the Pangea supercontinent and surrounding shelf seas and oceans. The interrelated succession of such changes is, however, insufficiently studied so far.

This work is dedicated to most important paleogeographic features and peculiarities in distribution of sedimentation settings. To define them, we compiled global lithological–paleogeographic maps for three time slices: late Sakmarian–early Artinskian of the Early Permian (Fig. 1), late Kazanian–early Toarcian of the Late Permian (Fig. 2), and Induan Age of the Early Triassic (Fig. 3). Reconstructions prepared for the international project "Pangea" (Scotese and Langford, 1995; Golonka *et al.*, 1994) global topographic schemes for every Permian age (Ziegler *et al.*, 1997, 1998) were used as the cartographic base for these maps. The maps summarize and systematize data published in works presented in the reference list. Difficulties accompanying interregional correlation of Permian deposits are well known. Several slightly different schemes of such correlation were proposed recently. In this work, we used the stratigraphic scheme proposed for the project "Pangea" (Ross *et al.*, 1994). Maps presented here are used to reconstruct most significant paleogeographic and depositional transformations that occurred on the Earth during the Permian and Early Triassic. Owing to known differences in correlation of events, the transformation ages estimated here are conditional to a certain extent, but this does not affect their essence and succession.

2. MAIN FEATURES OF THE PERMIAN–TRIASSIC PALEOGEOGRAPHY

Comparing the lithological–paleogeographic maps, one can see that disposition of global paleogeographic features was almost unchangeable during the considered time period. The Earth retained the longitudinal asymmetry: its oceanic hemisphere was occupied by the Panthalassa, and continental hemisphere congregated the huge Pangea continent, Paleotethys and Neot-

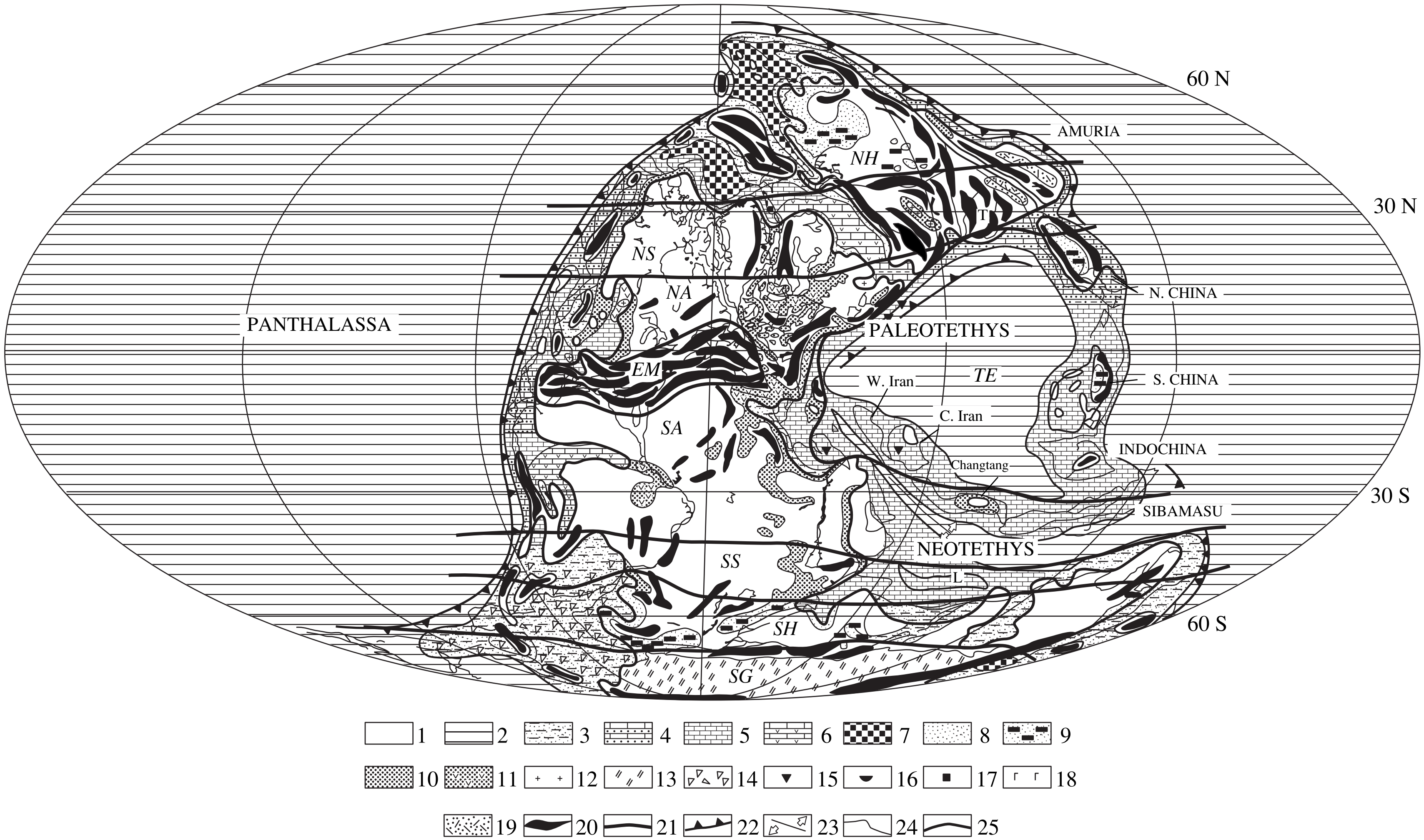

Fig. 1. Lithological–paleogeographic map for the Late Sakmarian–early Artinskian time of the Early Permian: (1) land; (2) oceans; (3) shelf seas with terrigenous sedimentation; (4) shelf seas with terrigenous–carbonate sedimentation; (5) carbonate platforms; (6) evaporite–carbonate platforms; (7) anoxic basins with black shale sedimentation; (8) inland and coastal alluvial and alluvial–lacustrine basins; (9) coal-bearing basins; (10) inland and coastal basins of arid zones that accumulated alluvial, eolian, and lacustrine red beds; (11) inland and coastal basins with alluvial, lacustrine, and sabkha gypsum-bearing red beds; (12) saliferous basins; (13) areas of predominant continental glacial sedimentation; (14) areas of predominant marine glacial sedimentation; (15) laterites, bauxites; (16) kaolin clays, kaolin-bearing sediments; (17) iron ores; (18) terrestrial basalts; (19) volcanogenic–sedimentary deposits; (20) orogenic structures; (21) boundaries of sedimentation–climatic belts; (22) subduction zones; (23) axes of spreading; (24) modern shore line; (25) ancient shore line. Belt types: (EM) equatorial–mountainous; (NA) northern arid, evaporite; (NS) northern semiarid; (NH) northern humid, coal-bearing; (SA) southern arid, evaporite; (SS) southern semiarid; (SH) southern humid, coal-bearing; (SG) southern glacial; (TE) tropical–equatorial coal- and bauxite-bearing; (T) Tarim and (L) Lhasa blocks.

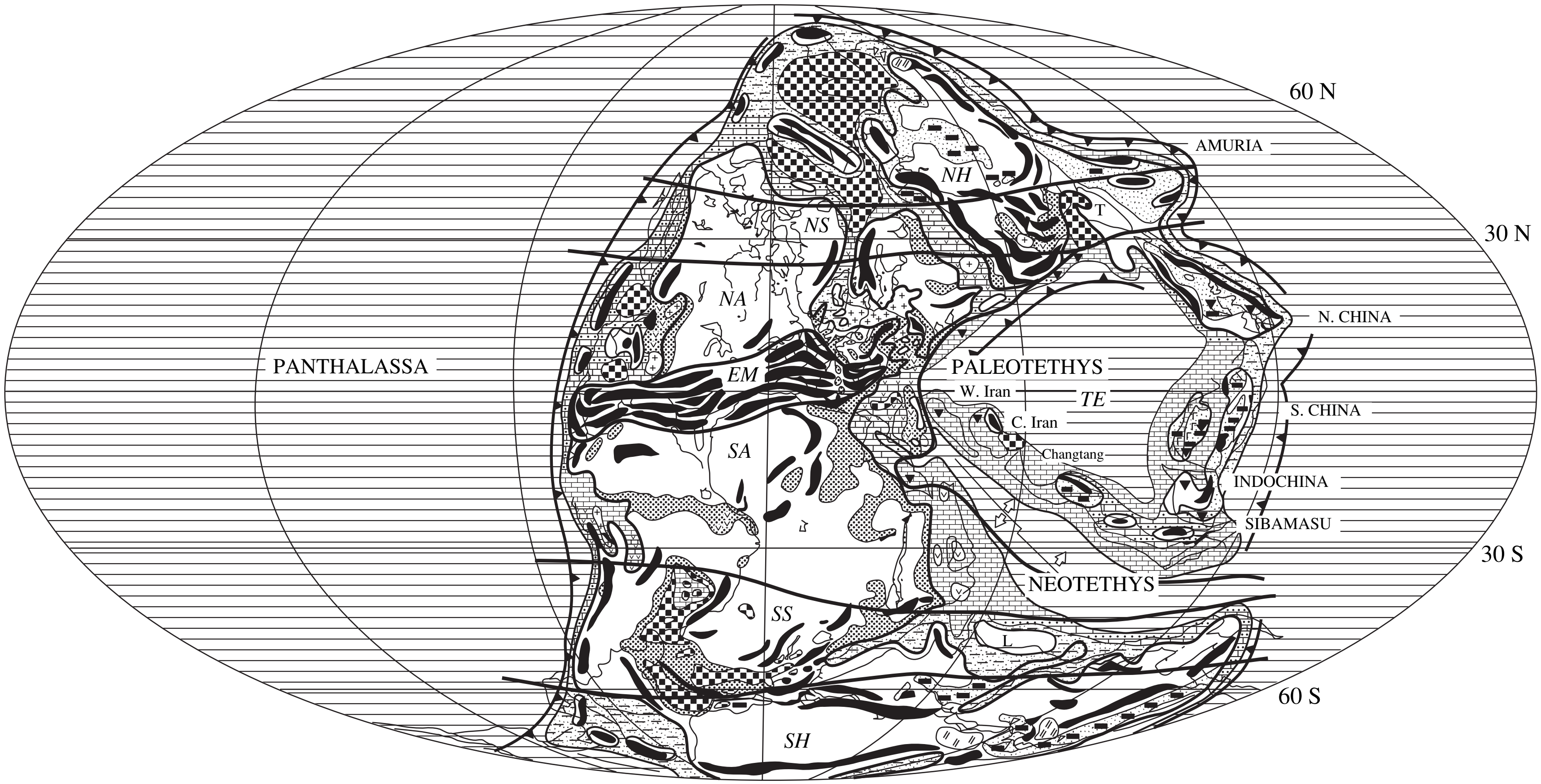

Fig. 2. Lithological–paleogeographic map for the Late Kazanian–early Tatarian time of the Late Permian (symbols as in Fig. 1).

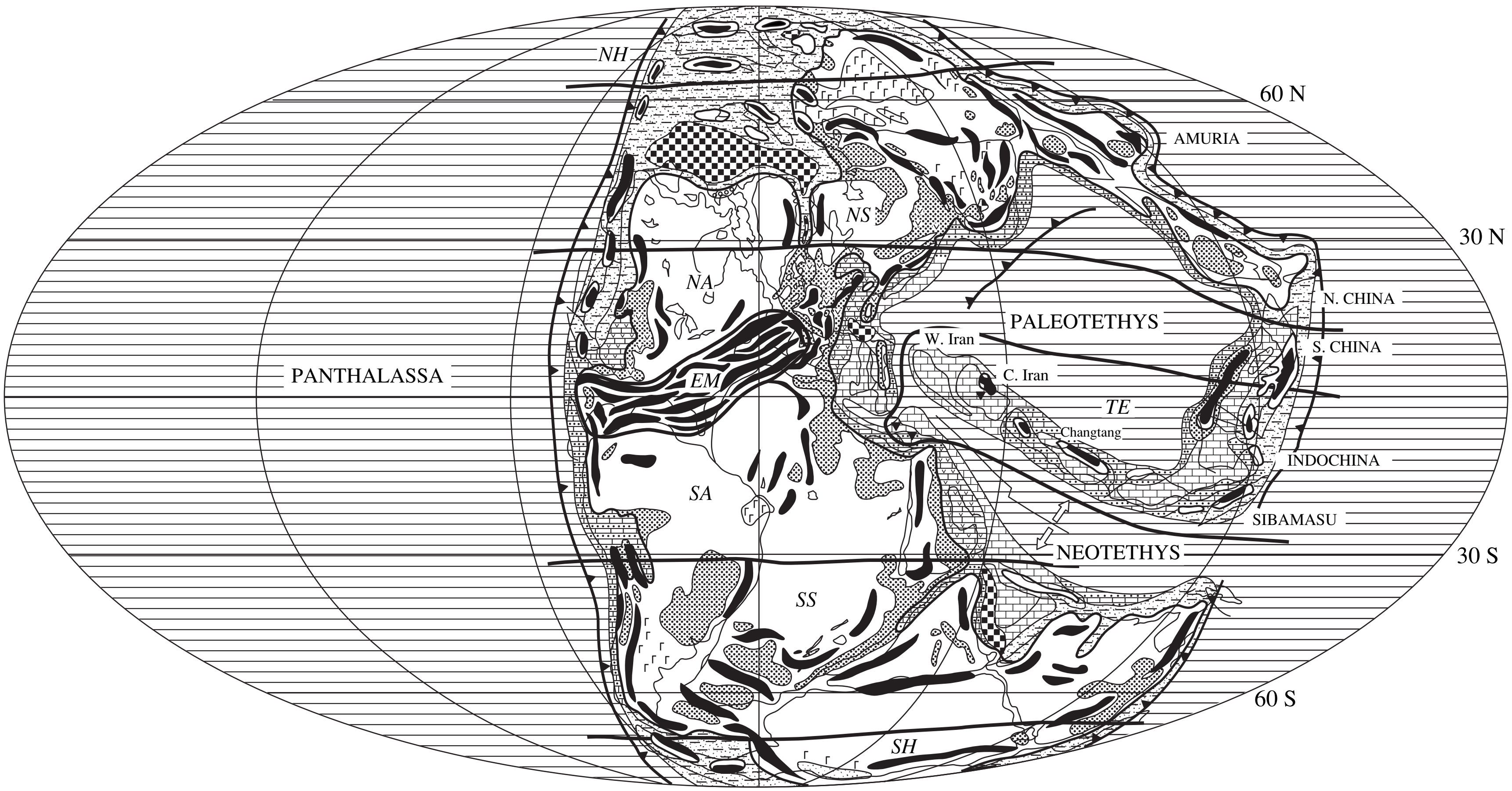

Fig. 3. Lithological–paleogeographic map for the Induan Age of the Early Triassic (symbols as in Fig. 1).

ethys oceans, Cathaysian and Cimmerian systems of microcontinents (Metcalfe, 1994; Golonka *et al.*, 1994; Scotese and Langford, 1995; Ziegler *et al.*, 1997, 1998). The Pangea that amalgamated the Laurasian continents (North America, Baltia, Siberia, and Kazakhstan) and Gondwanan (South America, Africa, Hindustan Peninsula, Australia, and Antarctica) groups of continents extended, as a single supercontinent, from the south pole to 75–85°N crossing all latitudinal belts. The Cathaysian system comprising the North and South Chinese microcontinents extended almost in the longitudinal direction bordering the Paleotethys in the east (Scotese and Langford, 1995). The Cimmerian system separated the Neotethys from the Paleotethys and included microcontinents of Western and Central Iran, North Tibet (Qiangtang), and Burma–Malaysia (Sibamasu) (Scotese and Langford, 1995). The period in question was marked by a slight northward drift of Pangea (by 5–10°) (Scotese and Langford, 1995; Ziegler, 1997, 1998). The most intense northward drift was characteristic of the Cimmerian and Cathaysian systems of microcontinents in connection with the Neotethys extension (Scotese and Langford, 1995; Ziegler *et al.*, 1998).

The principal stability in disposition of main planetary paleogeographic structures suggests that biosphere transformations in the Permian and initial Triassic geological history were probably caused by durable and successive changes in Pangea that slowly transformed its interrelation with surrounding seas and oceans, rather than by rapid and drastic paleotectonic and paleogeographic events. Accordingly, a special attention should be paid to paleogeographic evolution of the supercontinent Pangea and to changes in sedimentation settings within its territory and in adjacent seas.

Main orographic features of Pangea during the late Sakmarian–early Artinskian time were as follows. Its central near-equatorial part was occupied by a tremendous collision system of Central Pangea mountains (Scotese and Langford, 1995; Ziegler *et al.*, 1997), which crossed the supercontinent in the latitudinal direction dividing it into two halves: the northern (Laurasian) and southern (Gondwanan) parts. The inner continental areas of the Laurasian and Gondwanan segments of the supercontinent hosted collision-related mountain systems and belts, remnant Precambrian and Paleozoic mountains, arched uplifts, and plateau-like rises (Ziegler *et al.*, 1997). The collision systems of mountains were located in the Kazakhstan–Angara region of Laurasia. The Byrranga, Urals, and Kyzylkum mountain belts extended along the western and southern margins of the Laurasian part; the Sayan, Altai, and Verkhoyansk systems stretched along its eastern periphery, and the Yenisei–Zaisan mountain system occupied the central part (Ziegler *et al.*, 1997). The remnant and intraplate mountain systems were represented by the Antler, Grenville, and Ancestral mountains, Front Range, and Uncompahgre uplift in North America and by Scandinavian Mountains in western Baltia (Ziegler *et al.*, 1997). Large regions in the Laurasian part of Pangea were occupied by plateau-like rises such as the Baikalian and Patom uplands of the Angaraland and the Oslo highland in Baltia. The inner and remnant mountain systems in the Gondwanan part of Pangea included the Asuncion Mountains located along the western border of the Parana Basin in South America, the Espinhaco mountains in eastern South America, the Mauritania belt and the Ougarta and Iforas mountains in northwestern Africa, the Lomagundi, Muchinga, Makutu, Mikumi, Atacora, Mayombe, Mitumba, and Windhoek mountain belts in the eastern, central, and southern parts of Africa, and the Lachland Highland and the MacDonnel, Flinders, Hamersley and Musgrave ranges in Australia (Ziegler *et al.*, 1997). There were also the Ahaggar and Ennedi plateau-like and arched uplifts in Africa and the Great Plains of Australia. The mountainous structures of the Andean type stretched along the western extremity of South America (Andes), southern margin of Antarctica, and in eastern Australia (New England) (Scotese and Langford, 1995; Ziegler *et al.*, 1997).

It can be assumed that, in their orographic structure, many inner regions of Pangea were probably similar, to a significant extent, to modern drainless uplands, such as the Central African and Botswana plains and plateaus of the Central Iran, Gobi, Central Atlas, or High Plains of North America, West Australian plateau and others. These paleogeographic peculiarities predetermined a wide development of spacious blind drainage areas in both the arid and humid climatic zones.

Simultaneously, the shelf, marginal, and inland seas still occupied spacious areas of Pangea during the late Sakmarian–early Artinskian. In the northern periphery of Pangea, there were the Verkhoyansk–Chukchi, Sverdrup, and Barents marginal seas. The Midcontinent, Midland, Delaware, Williston, and other inland seas extended far inside western North America. They were separated from the western shelf seas by the Uncompahgre, Pedernal, and Diablo uplifts and islands. The large East European inland sea was located almost in the center of the Laurasian half of Pangea. It served as a seaway for the intermittent water exchange between the Paleotethys and Arctic basins. The entire periphery of the Gondwanan part, represented a spacious marginal sea located between South America, South Africa, and Antarctica. The Peru–Bolivian and sub-Andean sea gulfs were located in the west of Gondwana, and the Mozambique–Madagascar and West Australian seas of the same type occupied the eastern areas. These inland seas and gulfs divided the Pangea into several independent land areas of the Kazakhstan–Angaraland region, Laurentia (North America + West Europe), Western Gondwana, and Eastern Gondwana. Worthy of mention are also large marginal seas that were located along the eastern periphery of Pangea and occupied Italy and Dinarides in Europe, northern areas of Africa, almost entire northern half of the Arabian Peninsula, and the northern margin of Hindustan.

The main trend in the paleogeographic evolution of Pangea during the Late Permian–Early Triassic (Induan Age) was controlled by its steady rising. Geological records evidence the gradual sea retreat from inner areas and the reduction of marginal sea areas along the northern and southern peripheries of Pangea. Par example, the East European inland sea completely disappeared in Laurasia by the end of the Permian. Marginal seas located in northwestern and western Angaraland, as well as inland seas of western North America, reduced. Significantly reduced in dimensions were also marginal sea that occupied the southeastern periphery of the Gondwanan part of Pangea. The regression of inner and marginal seas was most likely conditioned by the Pangea uplift rather than by the global sea-level fall, as some researchers assumed (Schopf, 1974; Parrish, 1995; and others). This is evident from the fact that dimensions and positions of shelf zones around Pangea remained almost unchanged. Only marginal seas experienced reducing, whereas inland seas became replaced by basins of internal drainage. Thus, in the second half of the Late Permian and initial Early Triassic, Pangea represented a huge highly uplifted continent, the western, southern, and northeastern margins of which were occupied by mountain systems separating inner zones from shelf seas and oceans. The central part of Pangea hosted plateau-like rises and intraplate mountain belts, between which there were spacious drainless lowlands with meandering rivers and lake systems.

3. PERMIAN AND EARLY TRIASSIC SEDIMENTATION SETTINGS

The presented lithological–paleogeographic maps (Figs. 1–3) demonstrate sufficiently well the spatial distribution of sedimentation settings of the considered time period. They provide an opportunity to outline belts of arid, humid, and glacial settings and to clarify global changes in paleogeography, sedimentation, and climate. Working on the maps, we tend to define spacious and widespread lithological–facies zones of the same type within shelf seas and inland areas. Arid sedimentation settings represent lithological–paleogeographic zones of four types: (1) inland and coastal zones of alluvial, eolian, and lacustrine red beds; (2) inland and coastal zones of alluvial, lacustrine, and sabkha gypsum-bearing red beds; (3) salt-accumulating basins; and (4) evaporite–carbonate platforms. The humid sedimentation settings are divided into two groups: (1) inland and coastal areas of alluvial and alluvial–lacustrine sedimentation; and (2) coal-bearing basins. Areas of glacial sedimentation are subdivided into zones of predominantly terrestrial or marine settings. Zones of terrigenous and terrigenous–carbonate sedimentation, black-shale accumulation (anoxic environments), and carbonate platforms are distinguished among shelf settings. In addition, the most important paleoclimatic indicators, e.g., laterite, bauxite, kaolin, and iron ores deposits are shown in the maps out of scale.

3.1. Arid Sedimentation Settings

Huge dimensions of Pangea and extended mountain systems along its margins, as well as gradual regression of epicontinental seas and global warming owing to deglaciation, resulted eventually in the increased aridity of inland areas in low and middle latitudes and in displacement of semiarid belts toward poles (Robinson, 1973; Parrish *et al.*, 1986; Parrish, 1993, 1995; Barron and Fawcett, 1995). These changes in the spatial distribution of arid and semiarid settings are well seen in the lithological–paleogeographic maps (Figs. 1–3).

Beginning from the initial Sakmarian time of the Early Permian to the Induan Age of the Early Triassic, all central areas of Pangea located north and south of the mountain system of Central Pangea represented settings of arid sedimentation that outline the northern and southern arid belts. The northern arid belt is well distinguishable in the southern Laurasian part owing to the wide development of eolian deposits, terrestrial and coastal sabkhas, evaporite and saliferous basins, intermittently draining river valleys (wadis), deserts with saline playa lakes, alluvial and alluvial–lacustrine red beds (Mckee *et al.*, 1967; Glennie, 1983; Zharkov, 1974; Merzlyakov, 1979; Peterson, 1980; Drong *et al.*, 1982; Ziegler, 1982; Clemmensen and Abrahamsen, 1983; Glennie and Buller, 1983; Mazzullo *et al.*, 1985; Kukhtinov, 1987a, 1987b; Frenzel *et al.*, 1988; Johnson *et al.*, 1988; Sneh, 1988; Afzali *et al.*, 1989; Smith, 1989; Gast, 1991; Kiersnowski *et al.*, 1995; Mazzullo, 1995; Wardlaw *et al.*, 1995). The belt stretched from the western periphery of North America (Midcontinent, Williston, Delaware, and other basin) to central southern areas of Laurasia (East European, Dnieper–Donets, and Central European basins) and farther to southern margins of the Kazakhstan–Angaraland continent (Chu–Sarysu basin). The southern arid belt is also outlined rather confidently in the northern half of the Gondwanan segment of Pangea. The western part of the southern belt hosted the Peru–Bolivian and sub-Andean saliferous basins. In the west and northwest of Argentina, eolian sediments are recorded (Limarino and Spalletti, 1986). In central and eastern inland areas, arid alluvial–lacustrine red beds, evaporites, and desert settings are registered in the Amazon, Parnaiba, Barreirinhas, Gabon, Mali–Nigeria, North Sahara, Murzuq, Kufrah, Abyad, and Mozambique basins (Vysotskii *et al.*, 1973; Blant, 1973; Nairn and Smithwick, 1976; Cahen *et al.*, 1984; Klitzsch, 1990; Kogbe and Burollet, 1990; Lefran and Guirand, 1990; Wycisk, 1990; Baud *et al.*, 1993; Broutin *et al.*, 1995; Tankard *et al.*, 1995; Salem, 1996). The eastern periphery of the southern arid belt was occupied by a wide zone of carbonate–evaporite plateaus, coastal and continental sabkhas, shelf seas, saliferous and evaporite basins of the flooded and in-island types. The zone comprised salt basins of the North Italy, Dinarides, Mecsek, Moesian, Arabian, and other regions (Murris, 1980; Berberian and King, 1981; Zharkov, 1981; Sharief, 1981, 1983; Cassins *et al.*, 1992; Husseini, 1992; Ustaömer and Robertson, 1993; Alsharhan and Nairn, 1995).

The northern and southern arid belts were divided by the mountain system of Central Pangea. The system was likely characterized by the Himalayan-type morphology and, probably, by the vertical succession of climatic zones from the piedmont deserts to mountainous–steppe and mountainous–meadow settings in the intermediate and upper belts. In intermontane areas of the eastern system margin, which represented basins of internal drainage with branching and meandering rivers and lakes, sedimentation occurred under semiarid conditions (Cassins *et al.*, 1992, 1995; Ori, 1988; Châteauneuf and Farjanel, 1989). Similar semiarid environments could locally exist also along the western periphery of the mountain system.

Beginning from the second half of the Early Permian, spatial distribution of arid belts peculiarly changed. The South Gondwanan arid belt moved to the north. During the late Sakmarian–early Artinskian time, it was located between 10–15° and 40–45°S, whereas in the Induan Age of the Early Triassic, its northern boundary was near equator, and the southern one had position at 30°S. These changes were to a significant extent related to the Pangea drift northward. As for the northern, Laurasian arid belt, it also migrated northward and simultaneously widened to a significant extent owing to migration of its northern boundary toward the pole. In the Late Permian–Early Triassic, the Laurasian arid belt occupied the permanent position between 5–15° and 30°S.

Very important changes in sedimentation patterns occurred in the Early Triassic and likely affected both the northern and southern arid belts. Salt accumulation terminated in all inland and coastal basins of arid belts at that time. Evaporite sedimentation was characteristic either of alluvial and alluvial–lacustrine plains (sedimentation conditions of continental sabkhas, playa lakes, and takyrs) or, less commonly, of pericratonic evaporite–carbonate platforms. Branching river systems and zones of meandering rivers became widespread (McKee, 1954; Kukhtinov, 1987b; Movshovich, 1987; Klitzsch, 1990; Wycisk, 1990). The changes were probably caused by the enhanced monsoon character of the climate with more contrast seasonal and annual alternation of arid and humid periods (Kutzbach and Gallimore, 1989; Parrish, 1995). It can be stated therefore that sedimentation settings in arid belts became slightly more humid in the Early Triassic, when sedimentation environments in them resembled those of semiarid areas. Other important changes that occurred in arid belts of that time are as follows: inland basins of internal drainage that accumulated red beds, alluvial–lacustrine, evaporite, and desert facies became greater and more abundant, particularly in the western part of Pangea, in response to progressive rising of the supercontinent and retreat of inland seas from its central areas.

3.2. Semiarid Sedimentation Settings

The Permian–Early Triassic time was marked by extremely wide development of areas with insufficient seasonal moistening. During short humid seasons, these areas accumulated sediments of widespread meandering rivers, semiarid alluvial fans, deposits of intermittent flows, eolian sand dunes, soil carbonate concretions and calcretes, red beds, and variegated alluvial and alluvial–lacustrine facies. Coeval black shale sedimentation took place in open and closed lakes. Prolonged arid periods were responsible for deposition of continental sabkha-type and saliferous sediments. It is assumed (Parrish, 1995) that moistening patterns of this kind reflect the monsoon influence that strongly weakened in inland areas of Pangea protected from ocean by mountains.

Semiarid settings of Pangea recorded in southern and northern hemispheres outline respectively the southern and northern semiarid belts. The southern semiarid belt of the Late Permian time is detectable with a sufficient confidence in southern areas of Africa and South America. Sediments of alluvial plains, zones of meandering rivers, and flood plains with ordinary and playa lakes are registered in the Karroo, Tanzania, and some other regions (Kreuser *et al.*, 1990; Smith, 1990; Turner; 1990; Yemane and Kelts, 1990). Sedimentary basins accumulated here thick alluvial–lacustrine varicolored sediments of typical semiarid alluvial fans. In the southeastern part of the Karroo Basin, sedimentary sequence includes also soil carbonate calcrete-like concretions and playa-type gypsum-bearing beds (Turner, 1990). In southern Africa and southeastern South America, zones of meandering rivers surrounded large freshwater lakes with black shale sedimentation (Yemane and Kelts, 1990; Yemane, 1993, 1994; Franca *et al.*, 1995). In proximity to large lakes, local climate was moderately humid, with mean annual temperatures of about 10°C and moderate seasonal variations (Yemane, 1993). In other areas of the southern semiarid belt, the monsoon climate was characterized by alternation of arid and humid seasons and by insignificant precipitation (Smith, 1990).

Semiarid settings of the northern belt are rather certainty recognized in the Moscow syneclise and Volga–Urals region, which corresponded to the central part of the Laurasian segment of Pangea at that time. The complex association of alluvial, alluvial–lacustrine, subaqueous evaporite, terrestrial, and coastal sabkha sediments was formed here during the late Kazanian–early Toarcian (Ignat'ev, 1963, 1987; Kuleva, 1980; Strok, 1987; Tverdokhlebov, 1987a; Tverdokhlebov and Shminke, 1990; Lozovskii, 1998a; Lozovskii and Blom, 1998; Lozovskii and Zharkov, 1998). Prevalent were settings of changeable alluvial–lacustrine plains with seasonal sedimentation cycles, sedimentation breaks, soil and caliche formation, and gypsum accumulation in playas and sabkhas. Periodically widespread were branching river systems and probable zones of meandering rivers farther from the provenance. Features of the humid climate become well manifested by the end of the Permian (Ignat'ev, 1987; Strok, 1987; Tverdokhlebov, 1987a). Another area of the semiarid sedimentation located in the northern belt is recognizable in the southern Chu–Sarysu depression of the Kazakhstan–Angaraland continent. Terrestrial alluvial–lacustrine, floodplain, and sabkha sediments are represented here by red beds, variegated facies, and saliferous sequences with interbeds of the Na-Ca-sulfate composition (glauberite) indicating alternation of warm humid, arid, and cold seasons. Eolian sedimentation probably developed as well. The winter temperatures could vary from –5 to –15°C, whereas summer temperatures could be as high as +20–+30°C (Zherebtsova, 1977). In general, the northern semiarid belt of the early Late Permian time extended in between 25–30°N and 35–40°N.

In the terminal Late Permian–Early Triassic, the semiarid belts gradually widened in both the southern and northern hemispheres mainly owing to migration of high-latitude boundaries toward the relevant poles. The southern boundary of the southern semiarid belt was located in the Early Triassic at about 70–75°S. This belt covered the greater part of Australia, where semiarid settings of alluvial–lacustrine red beds are registered in the Bowen, Maryborough, Springfield, and Tasmania basins (Brown *et al.*, 1970; Veevers, 1984). The northern boundary of the northern semiarid belt was located in the Early Triassic between 65 and 70°N. It covered northern Cisuralian regions of the East European platform, and also southern and central areas of Angaraland. In the Pechora, Korotaikha, Bol'shaya Synya, Kos'yu–Rogovo, Kuznetsk, Gorlovo, Dzhungar basins, and in western areas of the Verkhoyansk basin of the belt, prevalent were alluvial and alluvial–lacustrine settings that accumulated red and varicolored deposits of semiarid sedimentation (Chelyshev, 1972; Dagys *et al.*, 1979; Kalantar, 1987; Neustroeva and Bogomazov, 1987; Lozovskii and Balabanov, 1988).The northern semiarid belt included also the Tarim, North Chinese, and Amuria microcontinents with red beds, fluvial and varicolored freshwater-lacustrine sediments widespread in the Kuqa, North Chinese, Khanka, and other basins (Kotlyar, 1984; Wang Hongzhen, 1985; Durante, 1998; Lozovskii, 1998b).

Thus, semiarid conditions spread in Pangea during the Late Permian–Early Triassic over spacious areas in middle and high latitudes of both hemispheres. The prevalent sedimentation settings allow their belts of their development to be referred to as "semiarid alluvial–lacustrine" ones. The northern semiarid belt of the Early Triassic time occupied areas between 30 and 65–70°N. The southern, comparatively wide semiarid belt stretched between 30 and 70–75°S. As is seen, both belts were located symmetrically relative to equator. By the beginning of the Triassic, sedimentation settings turned out to be more or less uniform over the entire Pangea supercontinent. This is evident from disappearance of cold belts and also from widened belts of semiarid alluvial–lacustrine sedimentation and a slight humidity increase in arid belts.

3.3. Humid Sedimentation Settings

Humid sedimentation settings and corresponding belts are distinguished, with some degree of certainty, only for the Permian time. They are recognizable by distribution patterns of coal-bearing basins and gray-colored alluvial, alluvial–lacustrine, boggy, and floodplain sediments. The available data allow the northern and southern Permian humid belts to be outlined in Pangea (Fig. 1, 2).

In the northern belt, continental and coastal humid settings were widespread in Angaraland and the northern Aral region, where the Tunguska, Kuznetsk, Gorlovo, Pechora, and other coal-bearing basins are known (*Atlas…*, 1968; Chelyshev, 1972; Bogomazov *et al.*, 1984; Meyen and Golubeva, 1984; Yuzvitskii *et al.*, 1984; Gurevich, 1987; Betekhtina *et al.*, 1988; Durante and Mogucheva, 1998). In the late Sakmarian–early Artinskian time, the humid coal-bearing belt covered the entire northern margin of Pangea northward of 30–40°N. In the late Kazanian–early Toarcian time, the belt occupied the same areas of Angaraland and Cis-Urals northward of 40–45°N. Its northern part is known to host glacial marine and seasonal ice sediments (Chumakov, 1994).

The southern humid belt of coal accumulation was located in the late Sakmarian–early Artinskian between 50–55 and 70–75°S. It included the Karroo, Ruhuhu, Luangua, and other basins in southern Africa, the Sokoto Basin in western Madagascar, the Damodar, Jharia, Mahanadi, Satpura. and other basins in the Hindustan, and the Bowen Basin in eastern Australia (Ahmad, 1964; Brown *et al.*, 1970; Kreuser and Semkiwa, 1987; Cook, 1990; Kreuser *et al.*, 1990; Smith, 1990; Turner, 1990; Mishra, 1991; Mitra, 1991; Langford, 1992). In the Late Permian, the belt covered almost the whole area of East Gondwana, including Australia and Antarctica, and was located south of 55–60°S. In Africa, humid environments were preserved only at its southern extremity, where the Jugella Ferry-Vryheid coal-bearing basin was located (Yemane and Kelts, 1990). In Hindustan, the belt included the above-mentioned coal-bearing basins. The entire southeastern part of Australia with Bowen, Denison, Sydney, Tasmania, Marree, and other coal-bearing basins was also a part of the belt (Langford, 1992). Some coal-bearing basins (Sydney, Tasmania, Marree) enclose glacial marine and continental sediments of the Late Permian (Kazanian) age (Caputo and Crowell, 1985; Langford, 1992; Crowell, 1995; Eyles *et al.*, 1998). This suggests that intermittent glaciers appeared in the southern humid coal-bearing belt.

During the period under consideration, other settings of humid sedimentation existed in the Cathaysian and Cimmerian microcontinents and in marginal coastal areas of Pangea. They outline the equatorial humid zone of coal- and bauxite-bearing deposits. Coal-bearing basins with alluvial, alluvial–lacustrine, alluvial–floodplain, and boggy sediments are known within this zone in the North Chinese, Amuria, Tarim, South Chinese, and Quingtang microcontinents and in the Quidam, Qinling, and Songpan-Ganzi terranes (Kotlyar, 1984; Lee, 1984, 1985a, 1986b; Ulmishek, 1984; Sheng *et al.*, 1985; Wang Hongzhen, 1985; Yang Zunyi *et al.*, 1986; Enos, 1995; Durante, 1998). Bauxite-bearing sequences are registered in the North Chinese, South Chinese, Indochina, and Western Iran microcontinents, as well as in the Caucasian and Pamirs margins of Pangea (Leven, 1984; Enos, 1995).

In the Early Triassic, coal accumulation ceased over the entire territory of Pangea, and in Cathaysian and Cimmerian microcontinents. The humid belts of the Early Triassic time are distinguishable only in circumpolar zones of northern and southern hemispheres, judging from gray-colored rocks prevailing in sequences of the marginal areas of Angaraland and Antarctica (Dagys *et al.*, 1979; Sadovnikov and Orlova, 1997; Retablak, 1999).

Thus, it can be postulated that the Late Permian–Early Triassic history was associated with several important changes in the humid belts. First, settings favorable for coal accumulation disappeared in the Early Triassic everywhere, including the humid belts as well. Second, the semiarid sedimentation gradually replaced the humid one, thus reducing the distribution area of the latter. Third, the humid belts migrated toward the poles. All these changes were apparently related to global warming and growing aridity of the climate.

3.4. Glacial Sedimentation Settings

As was mentioned, most significant biosphere transformations commenced, when Late Paleozoic glaciations degraded in the Sakmarian Age of the Early Permian.

Most researchers believe that the peak of the Early Permian glaciation was in the Asselian–Sakmarian time (Visser, 1996; Crowell, 1995). At that time, glacial sheets covered high and middle latitudes of South America, Africa (southern Arabia and Madagascar included), India, Tibet, and Australia. That glaciation likely affected also the Malacca–Burma terrane. Antarctica was apparently covered by ice completely. The glacial belt was sometimes as wide as 45–50°. In these continents, ice shields and mountainous glaciers left abundant exaration signs (striated glacier bed with all characteristic structures, trough valleys, fjords), basal tillites, fluvioglacial, lacustrine–glacial, and glacial marine sediments. The latter are particularly widespread. The glacial marine sediments, which accumulated under influence of shelf glaciers, snowmelt runoff, and icebergs, were more or less reworked by underwater colluviation. In the late Sakmarian–initial Artinskian, glaciers began to retreat everywhere, and the glacial belt significantly narrowed. Its northern boundary moved close to the Southern Polar Circle (Fig. 1).

The Karroo, Kalahari, and Karasburg basins of South Africa accumulated the thick Early Permian sequence of glacial sediments (the upper part of the Dwyka Group). Along periphery of these basins and on dividing rises, the sequence is composed of terrestrial glacial sediments, whereas in their central and southeastern parts, it is mainly represented by glacial marine deposits. The last signs of ice rafting in South Africa are registered in the lower part of the Prince Albert Formation corresponding to the base of the Ecca Group presumably of Artinskian age.

In Australia, the Early Permian glacial sediments are preserved in many sedimentary basins between Tasmania in the south and Bonaparte Bay in the north, where they are traceable from the western to eastern coasts of the continent. Many researchers believe that Permian glaciations commenced there during the Sakmarian Age and, reducing in dimensions, continued to exist with breaks during the Artinskian, Kungurian, and to Kazanian ages (Crowell and Frakes, 1971), or even till the Ufimian Age (Eyles *et al.*, 1998). Other authors estimated that glaciation in Australia lasted from the Asselian to early Sakmarian, and that no shield glaciation occurred there after this period (Dickins, 1996; Lindsay, 1997).

The most important glaciation centers were located in the western, central, and southern parts of Australia (Lindsay, 1997), and apparently in the Great Artesian Basin (Frakes, 1979). A thick ice sheet covered mountainous areas in the eastern coast of Australia. Structures observed in the glacial beds and tillites indicate that coastal glaciers of southern Australia moved from the southeast, i.e., from Antarctica, to the northwest (Lindsay, 1997; Bourman and Alley, 1999). The ice-rafted material occurring in southeastern Australia could also be delivered mostly from Antarctica (Eyles *et al.*, 1998).

In South America, Lower Permian glacial deposits are recorded in several basins located south of 10°S (modern coordinates). The largest of them is the Parana Basin of southern Brazil. Upper Paleozoic glacial sediments are distinguished here as the Itarare Group, the larger upper part of which is referred to the Lower Permian, although its age is debatable. Some researchers suggest the Asselian–Sakmarian or Asselian–Artinskian age of the group (Franca *et al.*, 1995), but others refer it to uppermost layers to the Kungurian Stage.

Similarly to Lower Permian glacial sequences of other continents, the Itarare Group shows repeated alternation of glacial marine and terrestrial facies that recorded glacial and interglacial sedimentation episodes, respectively. The latter were frequently accompanied by coal accumulation. South America hosted several glaciation centers. Glaciers advanced into the eastern part of the Parana Basin from the southeast, i.e., from South Africa, and into its western areas from the Asuncion arch of South America (Franca *et al.*, 1995). Glaciation in the Sergipe-Alagoas Basin originated from Equatorial Africa.

Signs of Early Permian glaciations are fairly abundant also in Hindustan. Prevalent here are continental glacial sediments preserved in many grabens, but in the northern structures framing the peninsula, they are represented by glacial marine facies. Tillites, boulder conglomerates, and other glacial sediments occur at the base of the Sakmarian–Artinskian Talchir Formation (Chandra, 1992). In many areas, the formation overlies the glacier bed with characteristic signs of ice exaration. All glaciation signs, such as striation in the glacier bed, orientation of elongated rock fragments, and other structural features of tillites indicate, with few exceptions, the glacier advancement from the south and southeast (Ahmad, 1981), i.e., from Antarctica and Australia.

In Antarctica, Upper Paleozoic glacial sediments are widespread in the Transantarctic Mountains; they are also known from areas bordering the Ronne shelf glacier and from some coastal areas of the Atlantic segment of East Antarctica. Continental and marine facies of glacial sediments fill several sedimentary basins here. They are estimated to be the Carboniferous and Early Permian in age (Isbell *et al.*, 1997). The modern ice cover hampers the detailed reconstruction of Late Paleozoic glaciation, but, taking into consideration the fact that glaciers from this continent advanced in the initial Early Permian as far as to the southern Africa, India, and southern Australia, it is clear that the entire East Antarctica was ice-covered at that time. This glaciation could not degrade before the Artinskian, because Antarctic glaciers reached southern areas of Africa (Visser, 1997) and India (Chandra, 1992) during this age. The Antarctic icebergs carried detrital material to southeastern Australia up to the Kungurian Age (Eyles *et al.*, 1998).

Unequivocal indications of Early Permian glaciations have not been discovered in the northern hemisphere so far. There are only single signs of presence of ice-rafted material and glacial to fluvioglacial sediments (Andrianov, 1966; Bobin, 1940), but their genesis is still debatable (Chumakov, 1994).

The late Kazanian–early Tatarian time was marked only by episodic glacial sedimentation in circumpolar humid belts of the southern and northern hemispheres. In the southern circumpolar belt, indications of ice-rafting (“dropstones”) are recorded in Kazanian and Tatarian sequences of the Sydney and Tasmania coal-bearing basins, in the Marree Basin, and in the northern Prince Charles Mountains of southeastern Australia (Corwell and Frakes, 1971; Crowell, 1995; Eyles *et al.*, 1998).

In the circumpolar areas of northern hemisphere, glacial and glacial marine sediments of the Late Kazanian–early Tatarian age are established amid marine facies in the Kolyma and Okhotsk massifs (Epshtein, 1973; Chumakov, 1994). They are probably present as well in the Verkhoyansk fold belt (Andrianov, 1966) and Omolon massif (Kashik *et al.*, 1990). Presence of glacial marine sediments indicates existence of land glaciers in some northeastern areas of Pangea during the Late Permian. These glaciers reached sometimes the sea level and formed icebergs and, probably, shelf ice fields (Chumakov, 1994). These geological data are well consistent with paleomagnetic measurements and geodynamic reconstructions, according to which the northern humid, episodically glaciated belt was located north of 60°N (Khramov *et al.*, 1982; Scotese and Langford, 1995; Parfenov *et al.*, 1999) or 70°N (Ziegler *et al.*, 1998).

The Lower Triassic glacial sediments have not been established either in the northern, or in southern circumpolar areas of Pangea. As was noted, circumpolar zones in the Early Triassic were warm and humid in both hemispheres of the Earth.

3.5. Marine Settings of Carbonate and Evaporite–Carbonate Sedimentation

The analysis of spatial distribution of marine sedimentation settings during the Late Permian–Early Triassic reveals a remarkable feature, namely the extremely wide development of carbonate facies in many shelf seas surrounding Pangea and also Cathaysian and Cimmerian microcontinents. They are known as well as in many inland seas of Laurasia and Gondwana that is particularly true of the late Kazanian–early Tatarian time (Fig. 2).

Carbonate accumulation along the western periphery of Pangea occurred almost in all shelf seas located between 45 and 50°N. Largest shallow carbonate platforms of the open ramp shelves and carbonate platforms of fringed shelves in inland seas are outlined in the Peru, Peru–Bolivian, and sub-Andean basins located along the western margin of the Gondwanan segment of Pangea (Franca *et al.*, 1995; Helwig, 1972; Sempere, 1995). The evaporite–carbonate platforms also formed in inland sea basins.

Another large area of carbonate sedimentation in western Pangea corresponded to the western margin of North America. This zone hosted an extended system of interrelated marginal seas, the Chihuahua, Marfa, Pedregosa, Central Arizona, Central Utah, Dry Mountain, Phosphoria, Central Wyoming, and other shallow-shelf basins with fringed platforms (Franzel *et al.*, 1988; Johnson *et al.*, 1988; Mazzullo, 1995; McKee *et al.*, 1967; Peterzson, 1980; Rascoe, 1988; Ross and Ross, 1995; Wardlaw *et al.*, 1995). In the west, the zone was separated from the open ocean by isolated islands (Antler and others) and by a chain of volcanic island arcs, whereas in the east, it was bounded by mountain systems (Uncompahre, Pedernal, Diablo, and others), which separated the zone from evaporite basins of the Midcontinent. In general, the entire system of carbonate shelf seas, island arcs, and dividing straits represented a transitional area between the open ocean in the west and inland evaporite basins in the east (Zharkov, 1974, 1978, 1981). Several large basins advanced far into the continent, thus representing inland seas that resembled the gulfs connected via narrow straits with basins of the transitional area. One of such seas was the Delaware basin connected by the narrow Hovey Channel with the Marfa and Chiahuahua carbonate-accumulating seas. A peculiar succession of back-reef settings, pisolite carbonate shoals, terrigenous–carbonate tidal to lagoonal evaporite-accumulation zones, and coastal sabkha and lakes was situated around of the Delaware basin. Toward the central part of the basin, there is recorded a transition from massive reefs to the fore-reef slope and, then, to the deep (from 300 to 550 m) inner zone of the basin that accumulated terrigenous–calcareous sediments (Garber and Harris, 1993; Mazzullo, 1995). Later, the deep uncompensated zone accumulated sulfates and salts (Anderson and Dean, 1995; Lowenstein, 1988). All available data suggest that carbonate and evaporite sedimentation in all marginal and inland sea basins along the southwestern periphery of North America took place in warm waters of the arid tropical zone (Anderson and Dean, 1995).

Widespread carbonate sedimentation settings were characteristic of shelf seas adjacent to the northern margin of Pangea (Stemmerik, 1995; Beauchamp *et al.*, 1989; Trettin, 1989; Dixon and Dietrich, 1990; Davies and Nassichuk, 1991; Jensen and Sørensen, 1992; Koyi *et al.*, 1993; Beauchamp, 1995; Brevik *et al.*, 1995; Nassichuk, 1995; Stemmerik and Worsley, 1995). Carbonate platforms of fringed shelves of variable width and length are traceable along the southern slope of the Sverdrup Basin, in the eastern and, probably, western slopes of the Norwegian–Greenland Basin, and in the southeastern Barents Basin. Carbonate platforms of these basins substantially differed from each other. For instance, carbonates of the relatively narrow Sverdrup fringed shelf are characterized by the impoverished biota dominated by the so-called “bryozoan assemblage”. These carbonate rocks largely consist of low-Mg calcite and have the elevated glauconite content that points to deposition of shelf carbonates in temperate cold water environments (Beauchamp, 1993, 1995). The carbonate platform sequence of the Barents Sea includes massive brachiopod or bryozoan limestones and cherts with bioturbation signs, which accumulated in shallow, oxygen-rich and moderately cold waters (Stemmerik and Worsley, 1995). To the contrary, shelf carbonate platforms of the Norwegian–Greenland Basin are composed of bryozoan–algal mounds up to 70 m high and about 500 m wide. The mounds are surrounded by shallow oolitic and allogenic carbonates that are replaced in the coastal zone by sabkha, tidal, and lagoonal sediments, i.e., by calcareous marls, algal carbonates, nodular mosaic gypsum, and oolitic grainstones (Stemmerik, 1995). Evaporite–carbonate platforms of this kind formed under warm water conditions in the semiarid climatic zone.

Carbonate sedimentation settings along the eastern periphery of Pangea were particularly widespread. They were typical of all shelf seas located along the Gondwanan and Laurasian margins of Pangea between 45°S and 30°N. This belt of carbonate accumulation that stretched in the SE–NW and SW–NE directions in the southern and northern hemispheres, respectively, was located in the tropical–subtropical zone favorable for carbonate sedimentation and formation of evaporite–carbonate platforms in addition to ordinary carbonate platforms. One of the largest evaporite–carbonate platforms was located in the Arabian Peninsula. It occupied the almost entire northeastern half of the peninsula and stretched over a distance more than 4500 km, being from 1000 to 2500 km wide. This was a pericratonic fringed-shelf platform, where carbonates accumulated in littoral and intratidal settings and sabkhas of supratidal zone represented areas of sallow-water evaporite sedimentation (Murris, 1980; Berberian and King, 1981; Sharief, 1981, 1983; Husseini, 1992; Alsharhan and Nairn, 1995). In the northeast, the platform was bounded by extended reefal buildups fringing the deep-water continental slope (Murris, 1980; Alsharhan and Nairn, 1995). In the northwest, a relatively deep shelf, the site of carbonate–clayey sedimentation, separated the Arabian evaporite–carbonate platform from the Toros’lar carbonate platform that apparently was of the pericratonic ramp type and fringed the oceanic rift trough in the southwest. Farther to the northwest, there was the Tunisian carbonate shelf platform (Lys, 1988). The area northward of the trough was occupied by the spacious evaporite–carbonate platform that extended over areas of the Dolomitic and Carnic Alps, Dravt Ridge, Durmitor Massif, and High Karst zone of the Dinarides (Miljush, 1973; Buggisch *et al.*, 1976; Zharkov, 1981). Paleogeographic reconstructions suggest that the Italo–Dinarides platform contacted in the east the structural succession of land mass, Moesian Salt Basin, and the easternmost evaporite–carbonate platform (Zhukov *et al.*, 1976; Zharkov, 1981).

Within the shelf of the Paleotethys located along the southeastern margin of Laurasia, there was an extended Caucasian carbonate platform composed of various algal, bryozoan, brachiopod, sponge, coral, oolitic, and calcarenite limestones intercalated with abundant reefal buildups formed in the littoral and intratidal zones of the shallow shelf (Miklukho-Maklay and Miklukho-Maklay, 1966; Leven, 1993; Kotlyar *et al.*, 1984; Rostovtsev, 1984). The Caucasian carbonate platform was probably of the pericratonic fringed type, but it is not inconceivable that it represented several interrelated autonomous pericratonic carbonate platforms. The presumable Afghan–Pamirs pericratonic carbonate platform extended during the Late Permian along the northern shelf margin of the Paleotethys, where biohermal and organogenic–detrital carbonate buildups formed, whereas littoral and intratidal settings accumulated clayey limestones, clayey marls, and oolitic carbonates (Miklukho-Maklay, 1963; Leven, 1984; Dronov and Kafarskii, 1980; Leven *et al.*, 1989).

Another large region of the warm water carbonate sedimentation corresponded to the Cimmerian and Cathaysian microcontinents, large areas of which were, as was noted, occupied by shelf seas. Carbonate platforms were located along the entire southern and northern peripheries of the Cimmerian microcontinents stretching from Western Iran to Sibamasu. In the Cathaysian microcontinents, carbonate sedimentation was widespread along western margins and in some central areas. The territory of western and central Iran was occupied by a spacious carbonate platform, where intratidal and littoral zones of carbonate accumulation surrounded the relatively deep-water central areas of clayey–calcareous and black-shale sedimentation (Alsharhan and Nairn, 1995). In the northeast, the Iran platform likely joined the carbonate platform of the South Afghan median mass and, probably, the Suleiman–Kirtar region of southeastern Afghanistan. The continuous cover of the Upper Permian shelf facies, mainly of littoral and intratidal carbonate sediments, is traceable in the Zuru, Huspasrud, Hilmand, Tirin, Logar, and Argandab zones, as well as in the northwestern part of the Katawaz trough (Dronov and Kafarskii, 1980). Judging from the accepted version of plate reconstruction (Scotese and Langford, 1995), another carbonate platform of western and southern peripheral zones of the Quingtang microcontinent (northern Tibet) was located eastward of the likely integral Iran–South Afghan carbonate platform. In the west and south, it fringed the coastal zone of paralic and continental sedimentation that was located in the northern central area (Enos, 1995). As is assumed, this carbonate platform extended far to the east, over the southern and eastern Sibamasu margins (Hutchison, 1989). In the South Chinese microcontinent, the shallow carbonate platform originated in the extreme north and west of the craton; its eastern margin was occupied by buildups of the barrier reef (Sheng *et al.*, 1985; Enos, 1995). The greater part of the fringed-shelf carbonate platform stretched along the eastern and northeastern margins of the Amuria microcontinent of the Eastern Sikhote-Alin zone (Kotlyar, 1984).

Finally, we should mention here one of the largest inland basins of evaporite–carbonate sedimentation—the East European sea. Carbonate and local intermittent evaporite sedimentation continued in this basin up to the terminal Kazanian Age and, in places, up to the initial Tatarian Age (*Atlas…*, 1969; Zharkov, 1974). All the time, a spacious evaporite–carbonate platform that extended over the North Caspian syneclise, Volga–Urals region, and northern Moscow syneclise developed in the eastern part of the East European craton. Sedimentation in shallow-water shelf settings of the platform occurred under arid and semiarid conditions parallel to subordinate sulfate sedimentation in submarine settings and sabkhas. In addition, there were large saliferous basins in the northern margin of the North Caspian Basin and adjacent areas.

Thus, carbonate and evaporite sedimentation of the Late Permian time prevailed in shelf seas surrounding Pangea and the system of Cathaysian and Cimmerian microcontinents. Particularly widespread in the arid and semiarid climatic belts were warm-water sedimentation settings. All sea basins located around Pangea or in its interiors between 40–45°N and 40–50°S were of this kind. A nearly continuous belt of carbonate and evaporite–carbonate platforms stretched along the western periphery of Pangea within the tropical–subtropical zone. Another belt of evaporite–carbonate and carbonate platforms was even larger and extended along the eastern margin of Pangea from the Arabian Peninsula in the south to the Caucasian region and further to the north across the East European basin almost up to the Barents Sea. The third spacious belt of warm-water carbonate sedimentation united shelf seas of the Cimmerian and Cathaysian microcontinents. It was entirely located in the equatorial–tropical zone. Hiatuses in sedimentation frequent in some areas, karstification, and local bauxite accumulation (Western Iran, South Chinese carton) represent characteristic features of the platforms (Elon, 1995). During the Late Permian, carbonate accumulation occurred also in moderately cold environments. This type of sedimentation is recorded in the northern middle-latitude humid belt, where carbonate platforms developed on shelves of the Sverdrup and Barents basins. Peculiar Upper Permian carbonate sediments (mainly limestones), which were conditionally termed “Bivalvia reefs” (Ganelin, 1997), accumulated in the northern and southern temperate belts. In the north, they are recorded in the Vrangel Island, Omolon and Kolyma massifs, being known as well in southern New Zealand. These limestones bear taxonomically impoverished, but quantitatively abundant faunas, which lack thermophilic reef-builders and other dwellers of warm-water carbonate platforms. Such faunal assemblages reveal prevalence of pelecypods (sometimes very large) and brachiopods (Runnegar, 1984; Ganelin, 1997). In the Lower Permian sequences of Tasmania, similar limestones laterally grade into tillites (Rao, 1981).

The characteristic feature of all Permian shelf carbonate seas was very limited development of terrigenous sedimentation settings along their coastal zones. For instance, terrigenous sediments are registered only along the southwestern (coastal) margin of the Arabian evaporite–carbonate platform, in some local areas adjacent to small uplifts in Zagros, in the southeastern

Toros'lar region, and in the west of Arabian Peninsula (Alsharhan and Nairn, 1995). Narrow sandy shoals and beaches, or sabkha-type supratidal zones relatively small in extension are recognizable around many others carbonate and evaporite–carbonate platforms confined to western and eastern margins of Pangea. As is established in the Capitan shelf of the Delaware Basin, where zone of sabkha facies is up to several dozens kilometers wide (Garber and Harris, 1993; Mazzullo, 1995), even in the case of a wide zone with coastal sabkhas and intermittently drained lakes (playas), terrigenous sedimentation was of minor importance. Prevalent in this case were mosaic gypsum-bearing rocks alternating with dolomites. Along their coast-facing margins, the carbonate platforms of the Cimmerian terranes were bounded by very narrow zones of detrital sediments, as it is well exemplified by the Iran–South Afghan platform. Littoral terrigenous facies were relatively widespread only around the Khamdian Oldland of the South Chinese microcontinent at the beginning of the Tatarian Age (Enos, 1995). Similar, very narrow coastal zones of shelf terrigenous sedimentation are assumed to be the deposition sites of fine-grained sandstones in the fringe carbonate platforms of the Sverdrup and Barents basins (Beauchamp, 1995; Stemmerik and Worsley, 1995). Terrigenous sediments of coastal zones are recorded only along the Uralian margin of the East European Basin (Chuvashov, 1995).

The Early Triassic was marked by a significant reduction of shelf carbonate accumulation, particularly along the northern and western margins of Pangea, where spacious terrigenous shelves appeared. Nevertheless, the principal distribution patterns of carbonate platforms remained unchanged.

Limited development of coastal detrital sediments in most shelf carbonate seas indicates that carbonate and evaporite–carbonate platforms along lowland margins of Pangea and microcontinents of the Cathaysian and Cimmerian systems were located close to desert areas of the arid and semiarid climatic zones, as it is evident from presence of eolian sediments and dunes there. In the humid zones, carbonate platforms were located near boggy coasts with rare meandering rivers. Thus, the carbonate platforms blocked somehow the lowland margins of Pangea, and this indicates that prevalent in this supercontinent were paleogeographic environments of internal, but not external drainage.

3.6. Black Shale Sedimentation Settings

During the late Early Permian and in the Late Permian–Early Triassic, anoxic settings favorable for black shale sedimentation were characteristic of many shelf seas and inland lakes. They are recorded in all climatic belts.

Sufficiently wide development of cold to moderately cold anoxic setting of black shale sedimentation is established for shelf seas of the northern Pangea margin. In the late Sakmarian–early Artinskian and late Kazanian–early Tatarian periods, there was a zone of interrelated black shale basins that extended from the northern margin of Alaska to the eastern coast of the Barents Sea. The zone included the Sverdrup, Wandel, Björnöya, Hammerfest, Nordcap, Northeastern Spitsbergen, and, probably, northern Novaya Zemlya basins (Beauchamp, 1995; Stemmerik and Worsley, 1995). Deep inner areas of these basins were characterized by stagnant anoxic sedimentation environments. In the Sverdrup Basin, accumulation of spongolites occurred first parallel to prograding clinoform filling of deep-water zones and draining of separate areas of fringed carbonate platform that was characterized by successive accumulation of spicule-bearing clayey mud, interbeds and lenses of carbonates and glauconite-bearing sands, and then by predominant deposition of spongolites in the cold-water environments (Beauchamp, 1995). In the Wandel, Björnöya, Hammerfest, Nordcap, and Northeastern Spitsbergen, prevalent at that time was also sedimentation in the deep-water stagnant settings. The indicated basins accumulated siliceous spiculite shales with bioturbation signs in their central deep areas and dark gray clayey mud with thin interbeds of bioclastic carbonates (packstones and wackestones) in shallower shelf areas near adjacent carbonate platforms (Stemmerik and Worsley, 1995).

North of the Sverdrup–Barents zone of black shale sedimentation, there was the large Verkhoyansk–Chukchi basin with anoxic sedimentation environments. Boundaries of the latter can be established very arbitrarily on the basis of isolated and scattered distribution areas of black shales in the Kolyma–Chukchi and Verkhoyansk regions, and in islands of the Arctic Ocean (Epshtein, 1973; Ganelin, 1984a, 1984b, 1984c, 1984d, 1997; Ustritskii, 1984, 1993; Kashik *et al.*, 1990). In the Vrangel Is., the corresponding sequence is composed of laminated to varved micritic limestones emanating H_2S, black shales, and varicolored shales with siliceous–manganese concretions (Ganelin, 1997). In the southern Yana–Kolyma region, prevalent are black massive and laminated rock sequences of shale and siltstone–shale composition, which accumulated under calm stagnant hydrodynamic conditions in inner zones of the past sea basin (Epshtein, 1973). In general, cold-water and moderately cold-water deep settings of black shale sedimentation were characteristic of many basins located along the northern margin of Pangea in the middle-latitude humid and circumpolar, episodically glacial climatic belts.

Warm-water anoxic settings of black shale sedimentation are registered in several sea basins of subtropical and tropical belts. They are presumed for the Phosphoria, Chihuahua, and some other basins in the southwestern periphery of the North American margin of Pangea, where black shale sedimentation continued probably until the Kazanian Age (Sheldon *et al.*, 1967; Ettensohn, 1994; Woods *et al.*, 1991). Along the eastern periphery of Pangea, anoxic conditions were intermittently restored in many areas of the Arabian and Italian–Dinarides evaporite–carbonate platforms (Zharkov, 1981; Alsharhan and Nairn, 1995). In the southeastern Laurasian segment of Pangea and adjacent northwestern margin of the Tarim microcontinent, conditions favorable for black shale sedimentation existed for a long time in the North Tarim Basin (Enos, 1995). Another region with anoxic sedimentation environments corresponded to central and northeastern areas of the Central Iranian microcontinent, where black micritic bituminous limestones, black marls, and shales with cherty lenses and concretions accumulated in relatively deep-water inner areas of the carbonate platform (*Geological…*, 1977; Sharief, 1981, 1983; Husseini, 1992; Alsharhan and Nairn, 1995). It should be noted that all mentioned areas and basins of warm-water black shale sedimentation were confined to shallow shelf zones of seas located either close to evaporite basins or directly within evaporite–carbonate platforms.

Widespread, particularly since the Kazanian Age, became also lacustrine settings of black shale sedimentation. As was mentioned, an extended belt of spacious freshwater lakes stretched along the southern margin of Western Gondwana occupying significant areas of South America and South Africa (Padula, 1969; Yemane and Kelts, 1990; Yemane, 1993, 1994; Franca *et al.*, 1995). A large lake basin with black shale sedimentation presumably existed in the Turgai lowland of the southern Angaraland–Kazakhstan area of Laurasia (*Atlas…*, 1968).

In the Early Triassic, basins with black shale sedimentation were less common. In the northern margin of Pangea, there remained only the Sverdrup–Barents basin and another small basin in the Chukchi Peninsula (Dagys *et al.*, 1979; Trettin, 1989; Wignall and Hallam, 1992; Ettensohn, 1994; Wignall and Twitchett, 1996). The North Italian basin also continued to exist (Wignall and Hallam, 1992; Wignall and Twitchett, 1996). A new basin with black shale sedimentation appeared in the Cis-Himalayas zone of Hindustan (Kapoor and Tokuoka, 1985). In addition, the deep Early Triassic basin with black shale sedimentation extended along continental slopes of the Amuria and North Chinese microcontinents (Isazaki, 1994; Bragin, 2000).

It seems that several types of basins with anoxic black shale sedimentation can de distinguished. The first group includes the deep shelf basins located in marginal seas that surrounded Pangea. The Sverdrup–Barents and Kolyma–Chukchi basins can be referred to this type. Basins of the second group, such as the Cis-Himalayas and Central Iranian basins of black shale sedimentation, were situated in inner, deeper zones of carbonate platforms. Anoxic basins of the third type occupied an intermediate position between the saliferous and evaporite basins, on the one hand, and the shelf zones of carbonate accumulation, on the other. This type can be exemplified by the Phosphoria, Chihuahua, North Italian, North Tarim, and some other basins. Finally, inland lake basins with black shale sedimentation represent the fourth type.

It is noteworthy that many basins with black shale sedimentation were located close to zones of evaporite accumulation and to coeval or earlier saliferous basins of the Permian time. As is known (Zharkov *et al.*, 1979), deep areas of evaporite basins accumulate large amounts of highly concentrated brines, which later can enter the shelf zones and intermediate basins with groundwater runoff. The subsurface migration of heavy salty brines into subsided areas of shelves or deep basins of marginal seas resulted in water stratification and created stagnant conditions favorable for accumulation of sediments with the elevated content of organic matter. These conditions could stimulate formation of anoxic environments in the Sverdrup, Barents, Phosphoria, Chihuahua, North Italian, North Tarim, East European, Central European, Arabian, and other basins. The Kolyma–Chukchi, Barents, and Sverdrup basins were also characterized by the intermittent thermal stratification of water mass. Spatial distribution patterns of most basins with black shale sedimentation indicate that influx of oxygen-free waters from ocean was hardly responsible for appearance of stagnant environments in shelf and marginal seas, as is assumed by some researchers (Hallam, 1994; Wignall and Hallam, 1992, 1996; Knoll *et al.*, 1996; Wignall and Twitchet, 1996). The available data suggest that most basins with black shale sedimentation formed likely owing to either the influx of highly concentrated brines from evaporite basins that have been located along the Pangea margin, or to the groundwater runoff from the supercontinent. In general, it can be assumed that the groundwater runoff from Pangea during the entire period of its high stand in the Late Permian–Early Triassic could significantly influence surrounding shelf seas.

3.7. Marine Settings of Terrigenous Sedimentation

The remarkable feature of the Permian time is a relatively limited distribution of seas with terrigenous sedimentation around Pangea and also around Cimmerian and Cathaysian microcontinents. More or less significant shallow-water and, particularly, deep-water terrigenous sedimentation commenced only in some sea basins during the Kazanian Age of the Late Permian. It was usually confined either to circumpolar regions of the northern and southern hemispheres, or to eastern margins of Pangea and Cathaysian microcontinents.

The spacious region of marine terrigenous sedimentation extended over the North Alaskan, Yukon, Novaya Zemlya, Verkhoyansk–Okhotsk, and Kolyma–Omolon provinces in the northern Pangea margin (Ganelin, 1984a, 1984b, 1984c; Ustritskii, 1984, 1993; Chernyak, 1984; Kashik *et al.*, 1990; Beauchamp, 1995; Povysheva and Ustritskii, 1996; Sadovnikov and Orlova, 1997). The North Alaskan shallow-water terrigenous shelf was a deposition area of glauconite sandstones, limy argillites, and clayey limestones, which accumulated in coastal and inner shelf zones, sometimes under storm-wave influence (Beauchamp, 1995). The Novaya Zemlya terrigenous shelf of the Kazanian–Tatarian time accumulated an intricately alternating sequence of sandstones, siltstones, and argillites mainly in tidal and, probably, in proximal–distal zones of inner and outer shelves, where hydrodynamics was highly active. Coastal marine sediments of the Verkhoyansk region are frequently represented by sandstones. They likely accumulated in the intertidal zone under influence of tidal currents and storm waves. The terrigenous shelf of the Kolyma–Omolon region was characterized apparently by shallow-water tidal and wave hydrodynamics that is evident from alternation of glauconite sandstone, siliceous–clayey glauconite limestone, and coquina layers. As a rule, shallow terrigenous shelves of the North Alaskan, Yukon, Taimyr, and Kolyma–Omolon regions were relatively narrow in the Late Permian time and fringed deep inner zones of marginal sea basins with stagnant anoxic environments.

One more large region with shallow-water terrigenous shelf sedimentation is outlined in the northeastern periphery of Pangea along the Mongolian margin of Panthalassa (Shipulina, 1959; Zonenshain *et al.*, 1990). A thick sandstone–shale sequence of this region likely accumulated in the inner and outer shelf zones and, probably, on the continental slope in the course of turbidite sedimentation.

Two other regions with terrigenous marine sedimentation were confined to active continental margins. One of them was located in southeastern areas of the South Chinese microcontinent, and the second one corresponded to the New Zealand margin of the Gondwanan segment of Pangea. As is inferred (Hsü *et al.*, 1990), the South Chinese region represented during the Late Permian a deep piedmont trough that accumulated terrigenous turbidites and molasses, being situated along the northern margin of the Hunan microcontinent; the latter separated from the Yangtze platform by the opening Nanpanjiang ocean. It is assumed that the Late Permian rising and erosion of the Hunan active margin and adjacent island arcs gave rise to accumulation of flyschoid deposits and conglomerates in the deep-water trough. The narrow and shallow terrigenous shelf is also outlined along the southern passive margin of the Hunan microcontinent that faced the Gunanhai ocean.[1] The New Zealand region was characterized by accumulation of thick terrigenous, mainly green- or gray-colored sandy–clayey sediments. They accumulated on shallow shelves and in deep troughs along island arcs (Brown *et al.*, 1970; Steven and Speden, 1978; Veevers, 1984).

Relatively narrow terrigenous shelves are assumed to have been existed also along the western periphery of Pangea. An extended system of subduction-related orogenic structures and volcanic island arcs is outlined here. It stretched from the western margins of Canada and the United States (Nicola, Huntington, Hallavah, Klamath and other island arcs) to southern margins of South America and western Antarctica (the Andean subduction zone, Patagonia island arc, and others). These arcs could have shallow terrigenous shelves (Scotese and Langford, 1995). Nevertheless, no deep troughs with the thick Upper Permian turbidite sequence are established there.

It should be noted again that the terrigenous sea shelves of the Late Permian time were significantly smaller in dimensions and less widespread than shelf settings of carbonate sedimentation. This proportion remains valid in general for the Early Triassic time despite the larger dimensions of shelf areas with terrigenous sedimentation of that time, particularly in the northern and western margins of Pangea.

4. MAIN TRENDS IN EVOLUTION OF SEDIMENTATION–CLIMATIC ZONING

The analyzed spatial distribution of latitudinal belts with arid, semiarid, humid, and glacial sedimentation shows that the period from the late Sakmarian Age of the Early Permian to the Indian Age of the Early Triassic was marked by gradual widening of the arid and semiarid belts and by contrast lowering in the global sedimentation–climatic zoning.

The latitudinal asymmetry of climatic and sedimentation zones was most significant in the late Sakmarian–early Artinskian time. One glacial belt comprised at that time all circumpolar areas of the southern hemisphere located south of 60–75°S. The northern and southern coal-bearing belts were asymmetrically located relative to equator. The southern humid belt was comparatively narrow and stretched between 50–55 and 60–75°S. Some its areas experienced intermittent glaciations. The northern humid coal-bearing belt, unlike the southern one, was sufficiently wide, covering all Pangea regions located north of 30–40°N. Distinctly asymmetrical was location of semiarid and arid belts. These belts extended almost up to 50–55°S in the southern hemisphere and up to 30–40°N in the northern hemisphere.

In the late Kazanian–early Tatarian period, sedimentation belts of Pangea were more symmetrical in the southern and northern hemispheres. The equatorial mountainous belt retained its central position. Regions located to the south and north of it were occupied by arid and semiarid belts. Middle and high latitudes in both hemispheres corresponded to humid coal-bearing belts; small glaciation centers appeared episodically in their circumpolar areas. Simultaneously, some asymmetry in position of sedimentation–climatic belts continued to exist. For instance, the arid and semiarid belts were of different width and had different positions rel-

[1] In our lithological–paleogeographic maps based on geodynamic reconstructions by Scotese and Langford (1995), the Hunan microcontinent, as well as the Nanpanjiang and Hunanhai oceans are not indicated.

ative to equator. The arid belt of the southern hemisphere extended almost up to 35–40°S, whereas in the northern hemisphere, its counterpart was limited by latitude of 30°S. The southern semiarid belt, which was located between 35–40 and 50–60°S, was substantially wider and closer to the pole. The northern semiarid belt stretched between 30 and 40°N. A similar migration toward the pole is also registered for the southern humid coal-bearing belt. The asymmetric location of belts could be a consequence of the asymmetrical position of continental blocks and seas in circumpolar zones of the Earth relative to equator. A large continental block of East Pangea was located in southern high latitudes, whereas the land areas in northern circumpolar zone were much lesser and surrounded by seas. In the late Kazanian–early Tatarian time, the asymmetry was even more pronounced, because the large regions of the Laurasian segment of Pangea were still occupied by spacious inland and shelf seas.

The terminal Permian and initial Early Triassic were periods of less contrast latitudinal sedimentation zoning in Pangea, where the almost ideal symmetry relative to equator characterized positions of belts with arid, semiarid, and humid sedimentation. The belts of arid evaporite sedimentation were located between 30°S and 30°N. The semiarid alluvial–lacustrine belts extended between 30 and 65–70°N in the northern hemisphere and between 30 and 75°S in the southern hemisphere. The circumpolar zones of both hemispheres corresponded to the humid sedimentation belts. Prevalent in Pangea became settings of arid and semiarid sedimentation. As a result, basins of internal drainage with prevalent arid evaporite and semiarid alluvial–lacustrine sedimentation multiplied and became larger in the inner areas of Pangea by the end of the Permian.

5. GLOBAL EXCHANGE OF MATTER BETWEEN LAND AND OCEANS DURING PERMIAN–TRIASSIC BIOSPHERE TRANSFORMATIONS

Data discussed in previous sections indicate that most important paleogeographic changes of the Late Permian–Early Triassic time were the degradation of glacial belts and the gradual uplift of Pangea responsible for regression of inner and marginal seas. These events resulted in global warming, widened belts of arid and semiarid sedimentation settings, and increased areas and number of basins of internal drainage. Simultaneously, settings of humid sedimentation became less widespread that caused the gradual reduction of surficial runoff and displacement of terrigenous deposition centers from marginal to inner areas of Pangea.

In this connection, we should point to a significant disproportion between intensities of terrigenous sedimentation in terrestrial and marine settings, and also between volumes of carbonate and terrigenous sediments accumulated in the Late Permian. Even approximate estimates indicate that the intensity of continental terrigenous sedimentation in the second half of the Late Permian was about an order higher than that of marine terrigenous sedimentation. Simultaneously, the intensity of accumulation of carbonate sediments was 1.5–2 times higher than that of marine terrigenous deposits. In the Cretaceous, for instance, the proportion was opposite, and the carbonate/terrigenous sediment ratio was approximately 1 : 2 (Ronov, 1980). During the Pliocene, the volume of accumulated marine carbonates was 15 times lesser than that of marine terrigenous sediments (Ronov, 1980).

Nowadays, the intensity of biogenic carbonate sedimentation in shelf seas is also many times lesser than that of marine terrigenous sedimentation (Garrels and Mackenzee, 1974; Lisitsyn, 1974). All these proportions unambiguously indicate that the global exchange of matter between land and shelf seas of the Late Permian time was principally different. The unbalanced mineral exchange between the supercontinent and oceans was probably counterbalanced by the increased groundwater runoff into shelf zones of oceans.

The groundwater runoff volume can be estimated approximately in the following way. The current proportion between intensity of terrigenous marine and continental sedimentation is evidently close to the Pliocene one and could be about 1 : 2 (Ronov, 1980). As was mentioned, terrigenous sedimentation in the Late Permian marine settings was almost by an order less intense than in continental ones (proportion 1 : 10), i.e., 5 times slower than now. The present-day influx of different chemical components into oceans is approximately 237×10^{14} g/year; about 210×10^{14} g/year (88.6%) is contribution of the surface runoff, and only 4.3×10^{14} g/year (1.8%) characterize the groundwater runoff (Garrels and Mackenzee, 1974), i.e., the proportion is now about 50 : 1. In the terminal Permian, the contribution of groundwater runoff could be several times greater that now. As was shown for the Caspian and Aral seas, the groundwater runoff in modern arid and semiarid zones can be as great as 25–27% of the total continental runoff (Zektser *et al.*, 1984). Taking these assessments into consideration, one can assume that the groundwater runoff at the end of the Permian, when semiarid and arid settings prevailed in Pangea, could be at least by an order higher than the modern one. These approximate calculations provide only general idea about the groundwater runoff rate in the Late Permian. Nevertheless, they imply that the groundwater runoff from Pangea supercontinent into shelf seas was quite significant in the Late Permian.

The gradual reduction of the surface influx and increasing significance of the groundwater runoff should change substantially the global exchange of matter between Pangea and oceans. First of all, the total continental runoff should be of different chemical composition. The available data (Shvartsev, 1979; Meybeck, 1979; Savenko and Zakharova, 1997; Savenko, in press) indicate that it should be characterized by significantly higher Na, Mg, Ca, Cl, SO_4, and HCO_3 concentrations, but by the decreased P content. The influx of large volumes of anoxic underground Cl–Mg and Cl–Ca brines depleted in phosphorus and originated from arid and semiarid areas could significantly affect biota of shelf seas.

6. CONCLUSION

1. Paleogeographic evolution of Pangea and changes in sedimentation settings during the period from the late Sakmarian (Early Permian) to Induan (Early Triassic) ages progressed very slowly and depended on the general uplift of supercontinent, regression of inner and marginal seas, formation of marginal and inland collision orogenic belts, crustal arching, and plateau-like highlands, and plains. As a result, the drainless basins became widespread in Pangea. The surface runoff into surrounding shelf seas and oceans gradually decreased, while the inland basins of internal drainage became more widespread.

2. The Permian-Early Triassic global warming resulted in degradation of the glacial belt, widened arid and semiarid sedimentation zones, reduced humid areas, and decreased contrast of the sedimentation–climatic zoning.

3. At the beginning of the Early Triassic, salt accumulation and coal formation terminated throughout the entire Pangea supercontinent. This was likely caused by the greater influence of monsoons in low and middle latitudes of Pangea (Parrish, 1995) and also by the widening of semiarid sedimentation belts.

4. The Late Permian–Early Triassic time was characterized by extremely wide development of carbonate and evaporite–carbonate platforms. The platforms extended between 40–45°N and 40–50°S along the western and eastern margins of Pangea and around the Cimmerian and Cathaysian microcontinents. Conditions favorable for carbonate sedimentation in both the warm and moderately cold marine settings were controlled by the limited influx of detrital material from land into surrounding shelf seas, and this indicates that internal, but not external drainage prevailed on the supercontinent.

5. In the Late Permian, shelf seas with terrigenous sedimentation were of very restricted distribution. They were localized, as a rule, either along the northern and southern margins of Pangea in the middle and high latitude zones of the humid belts, or along the eastern periphery of the Cathaysian microcontinents in the tropical–equatorial humid belt. The insignificant development of terrigenous shelf sedimentation around Pangea additionally proves a minor significance of river runoff from the supercontinent.

6. In contrast, anoxic basins with black shale accumulation were very characteristic of the Permian–Early Triassic period. Three types of these basins are distinguishable: (1) deep-water shelf basins; (2) inner zones of carbonate and evaporite–carbonate platforms; and (3) inland lacustrine basins. Many shelf basins with black shale sedimentation were located near the coeval or older areas of evaporite accumulation, which gave rise to influx of heavy brines into shelf zones. As a consequence, stratification saline water mass resulted in appearance of stagnant environments favorable for accumulation of sediments with the elevated content of organic matter. Sufficiently wide development of such anoxic shelf basins around Pangea indicates that the underground brine flows played a significant role in the mineral exchange between the supercontinent and surrounding shelf seas.

7. The way for the Permian–Triassic biotic crisis was probably paved by the growing aridity of the post-glaciation Late Permian climate that substantially reduced the river discharge from Pangea and significantly increased the groundwater runoff. This drastically distorted the balance of mineral exchange between land and oceans and could cause the crisis because of reduction of primary production in oceans and total biomass of continental biota. The unavoidable results are the sharply decreased influx of nutrients from Pangea, the general biomass reduction, extinction of phytoplankton (Zezina, 1991), and subsequent destruction of the entire food chain in oceans.

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