Late Triassic Global Plate Tectonics

Jan Golonka [1]

Email jgolonka@agh.edu.pl

Ashton Embry [2]

Email ashton.embry@canada.ca

Michał Krobicki [~~3,4~~1,3]

Email: krobicki@geol.agh.edu.pl

[1] Faculty of Geology, Geophysics and Environmental Protection, AGH University of Science and Technology, Al. Mickiewicza 30, 30-059 Kraków, Poland

[2] Geological Survey of Canada, Calgary, AL, Canada

[3] Carpathian Branch, Polish Geological Institute – National Research Institute, Kraków, Poland

[4] AGH University of Science and Technology, Kraków, Poland

Abstract

The Late Triassic was the time of the Early Cimmerian and Indosinian orogenies that closed the Paleotethys Ocean, which occurred earlier in the Alpine-Carpathian-Mediterranean area, later in the Eastern Europe-Central Asia and latest in the South-East Asia. The Indochina Southeastern Asian and Qiangtang plates were sutured to South China. The new, large Chinese-SE Asian plate, including North and South China, Mongolia and eastern Cimmerian plates, was consolidated by the end Triassic, leaving open a large embayment of Panthalassa, known as Mongol-Okhotsk Ocean, between Mongolia and Laurasia,. The Uralian Orogeny, which sutured Siberia and Europe continued during Late Triassic times and was recorded in Novaya Zemlya. The onset of Pangaea break-up constitutes the main Late Triassic extensional event. Continental rifts originating then were filled with clastic deposits comprising mainly red beds. The pulling force of the north-dipping subduction along the northern margin of Neotethys caused drifting of a new set of plates from the passive Gondwana margin, dividing the Neotethys Ocean. Carbonate sedimentation dominated platforms on the Neotethys and Paleotethys margins as well as the Cimmerian microplates. Synorogenic turbidites and postorogenic molasses were associated with the Indosinian orogeny. The late stages of the Uralian orogeny in Timan-Pechora, Novaya Zemlya and eastern Barents regions filled the foreland basin with fine-grained, molasse sediments. Siliciclastics were common in the Siberia and Arctic regions. The widespread, large magnitude, base-level changes of the Late Triassic are interpreted as an expression of relatively rapid and substantial changes in the horizontal and vertical stress fields that affected the Pangaea supercontinent. Such stress changes may be due to abrupt changes in the speed and/or direction of plate movements, which episodically affected Pangaea.

Keywords

Paleogeography
Plate tectonics

Paleoenvironment
Paleolithofacies
Paleoclimate
Sea level changes

2.1. Introduction

The Triassic maps used here (Figs. 2.1, 2.2, 2.3, 2.4, 2.5, 2.6, 2.7 and 2.8) were derived from a series of global and regional Phanerozoic paleogeographic and plate tectonic maps which depicted present day coastlines, plate boundaries (sutures), selected transform faults, spreading centers, rifts, normal and thrust faults as well as paleoenvironment and lithofacies (Golonka 2000, 2002, 2007a, b, 2011; Golonka et al. 2003a, 2006a,b). Also included is a corrected and improved version of the Triassic maps previously presented (Golonka 2007a, b). The base maps, (past position of present day coastlines and plate boundaries) were generated by PLATES, PALEOMAP and GPLATES computer software (see Sect. 1.2). The definitions of mapped time slices were presented by Golonka and Kiessling (2002), however, recently the simple stratigraphic “Late Triassic” slice was used (Golonka 2007a, b). The name “Triassic” was derived from the German Trias defined by von Alberti (1834), referring to the division of the period into three stages: the Buntsandstein, Muschelkalk, and Keuper (see Köppen and Carter 2000; Feist-Burkhardt et al. 2008; Scheck-Wenderoth et al. 2008; McKie and Williams 2009 and references therein). This sequence is valid for Central Europe (Germany, Poland), but causes many problems when applied to other regions. The global Late Triassic (Ogg et al. 2016) is now divided into the Carnian, Norian and Rhaetian ages (Fig. 2.9). For the environment and facies assembly we used two units, applying the methods used for the Phanerozoic reefs map (Kiessling and Flügel 1999) and also presented by Golonka (2007a, b). The base maps (Figs. 2.1 and 2.2) depict the configuration of land masses, rifts, spreading centers and subduction and the beginning (Fig. 2.1) and end (Fig. 2.2) of the Late Triassic. The paleoenvironments and lithofacies (Figs. 2.3, 2.4, 2.5, 2.6, 2.7 and 2.8) represent the whole of the Late Triassic Epoch. They are posted on the 224 Ma base maps.

Fig. 2.1

Global plate tectonic map of Late Triassic at 224 Ma ago. Molweide Projection. (1) oceanic spreading center and transform faults, (2) subduction zone, (3) thrust fault, (4) normal fault, (5) transform fault

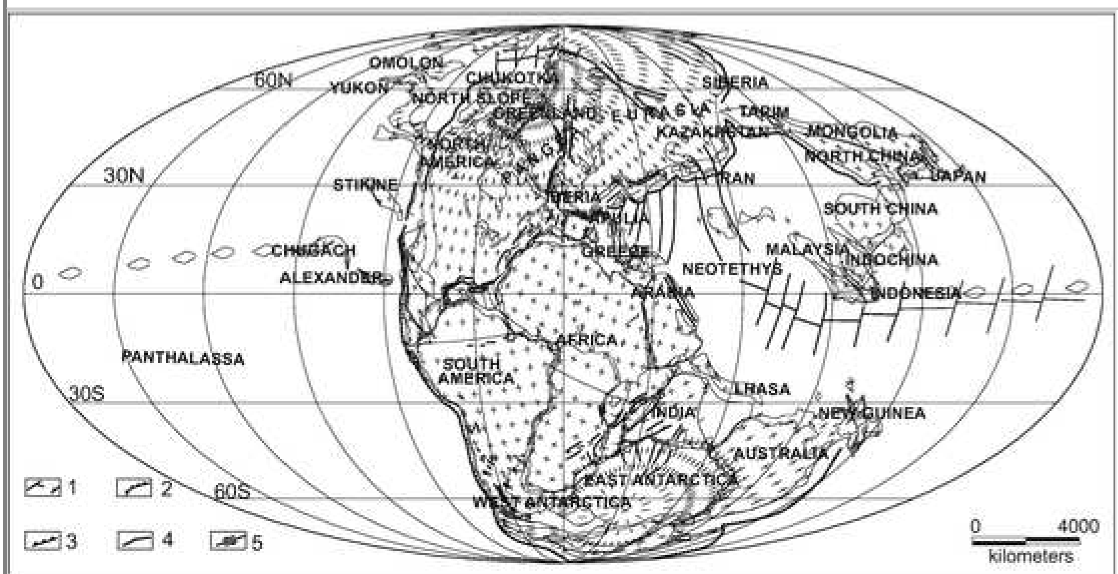


Fig. 2.2

Global plate tectonic map of Late Triassic at 200 Ma ago. Molweide Projection. (1) oceanic spreading center and transform faults, (2) subduction zone, (3) thrust fault, (4) normal fault, (5) transform fault

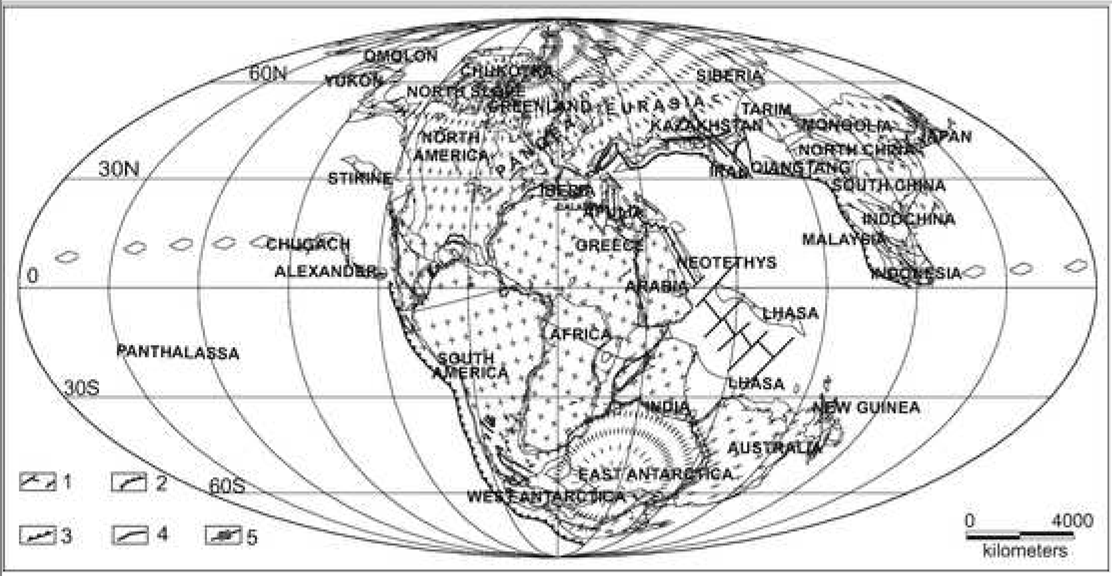


Fig. 2.3

Plate tectonic, paleoenvironment and lithofacies map of the western Tethys, future Central Atlantic and adjacent areas during Late Triassic time. Molweide Projection. Modified from Golonka (2007b)

Fig. 2.4

Explanations to Figs, 2.3, 2.4, 2.5, 2.7. Qualifiers: *B* bauxites/laterites, *C* coals, *E* evaporites, *F* flysch, *Fe* Iron, *G* glauconite, *M* marls, *O* oolites, *P* phosphates, *R* red beds, *Si* silica, *T* tillites, *V* volcanics

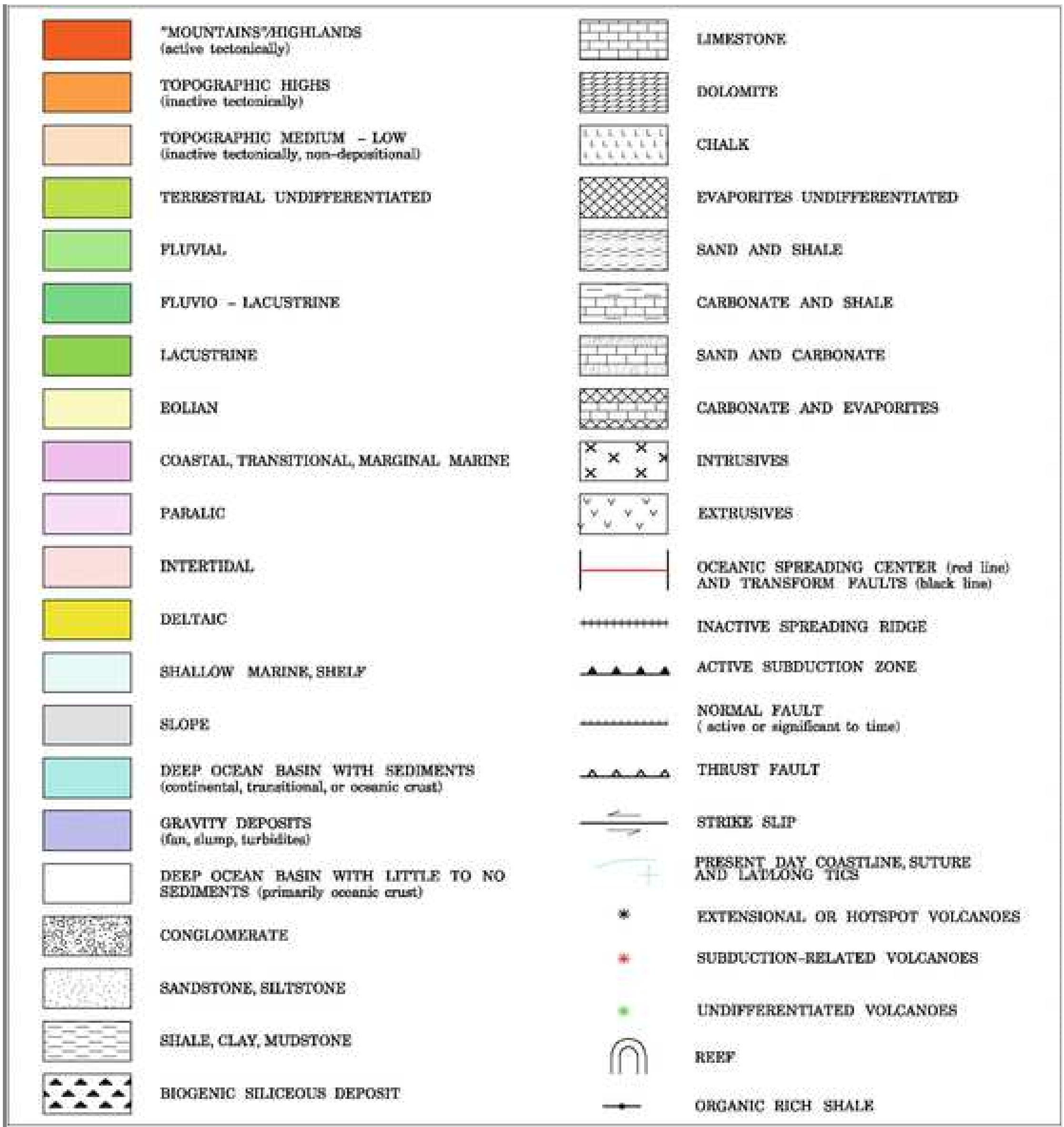


Fig. 2.5

Plate tectonic, paleoenvironment and lithofacies map of eastern Tethys and adjacent areas during Late Triassic time. Molweide Projection

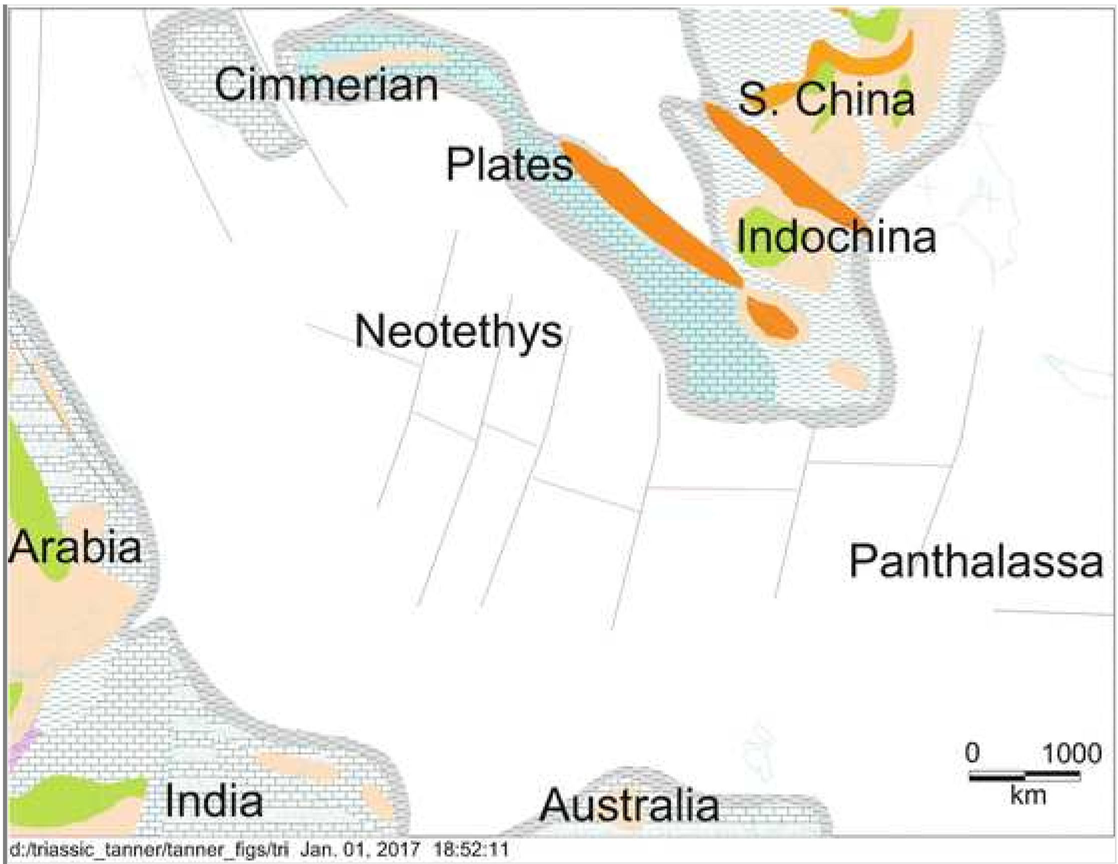


Fig. 2.6

Plate tectonic, paleoenvironment and lithofacies map of the Paleotethys, Chinese plates and adjacent areas during Late Triassic time. Molweide Projection. Modified from Golonka (2007b)

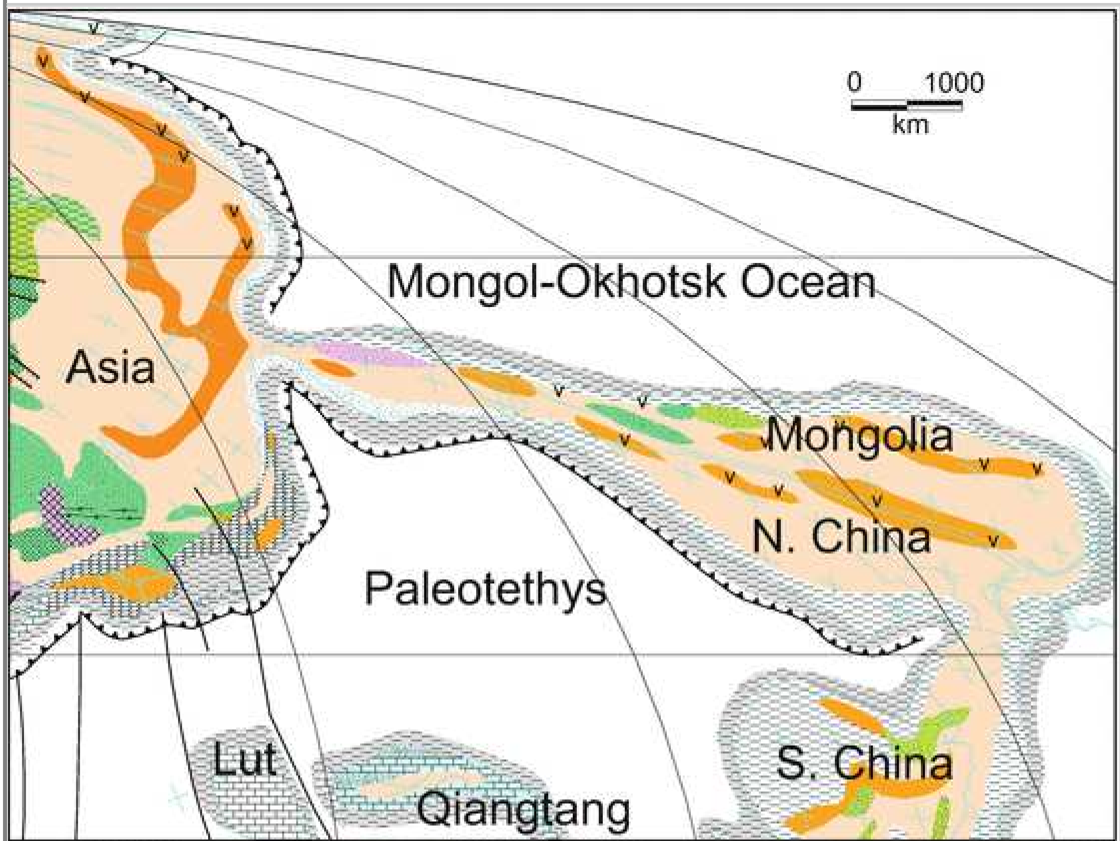


Fig. 2.7

Plate tectonic, paleoenvironment and lithofacies map of the western Gondwana and adjacent areas during Late Triassic time. Molweide Projection. Modified from Golonka (2007b)

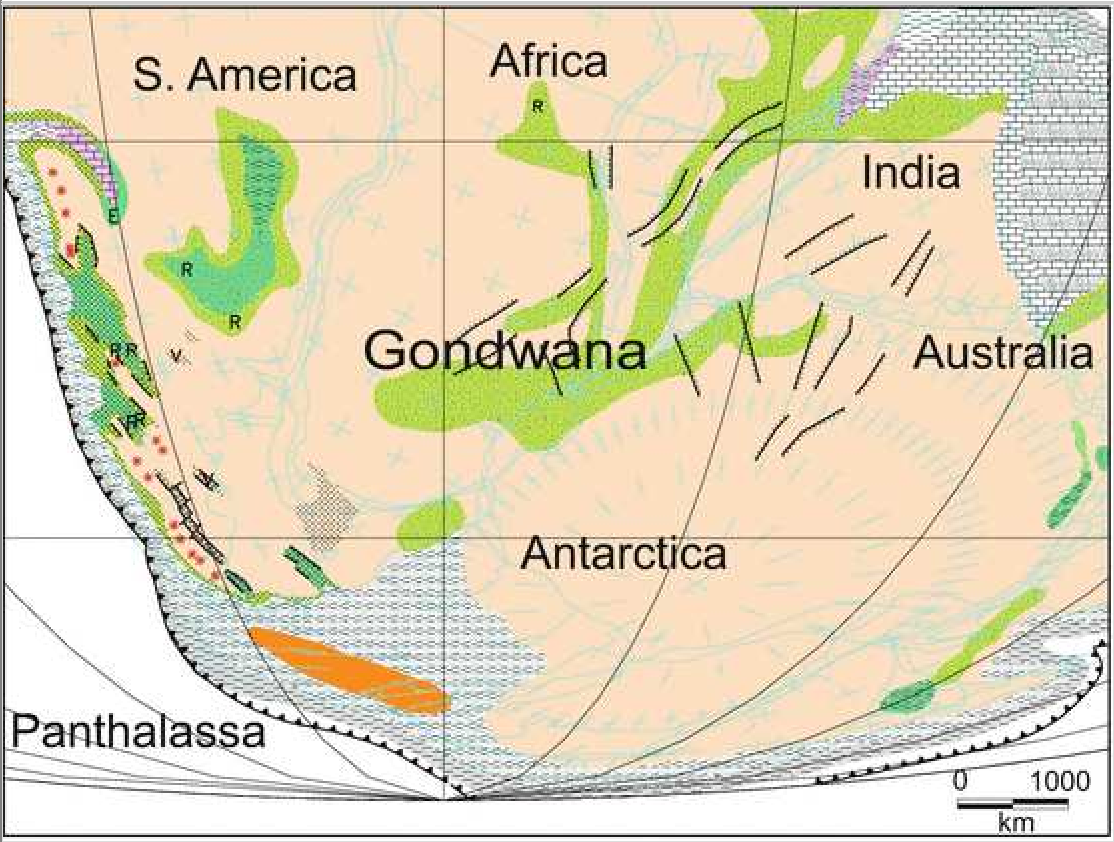


Fig. 2.8

Plate tectonic, paleoenvironment and lithofacies map of the Arctic during Late Triassic time. Stereographic polar. Projection Modified from Golonka (2011)

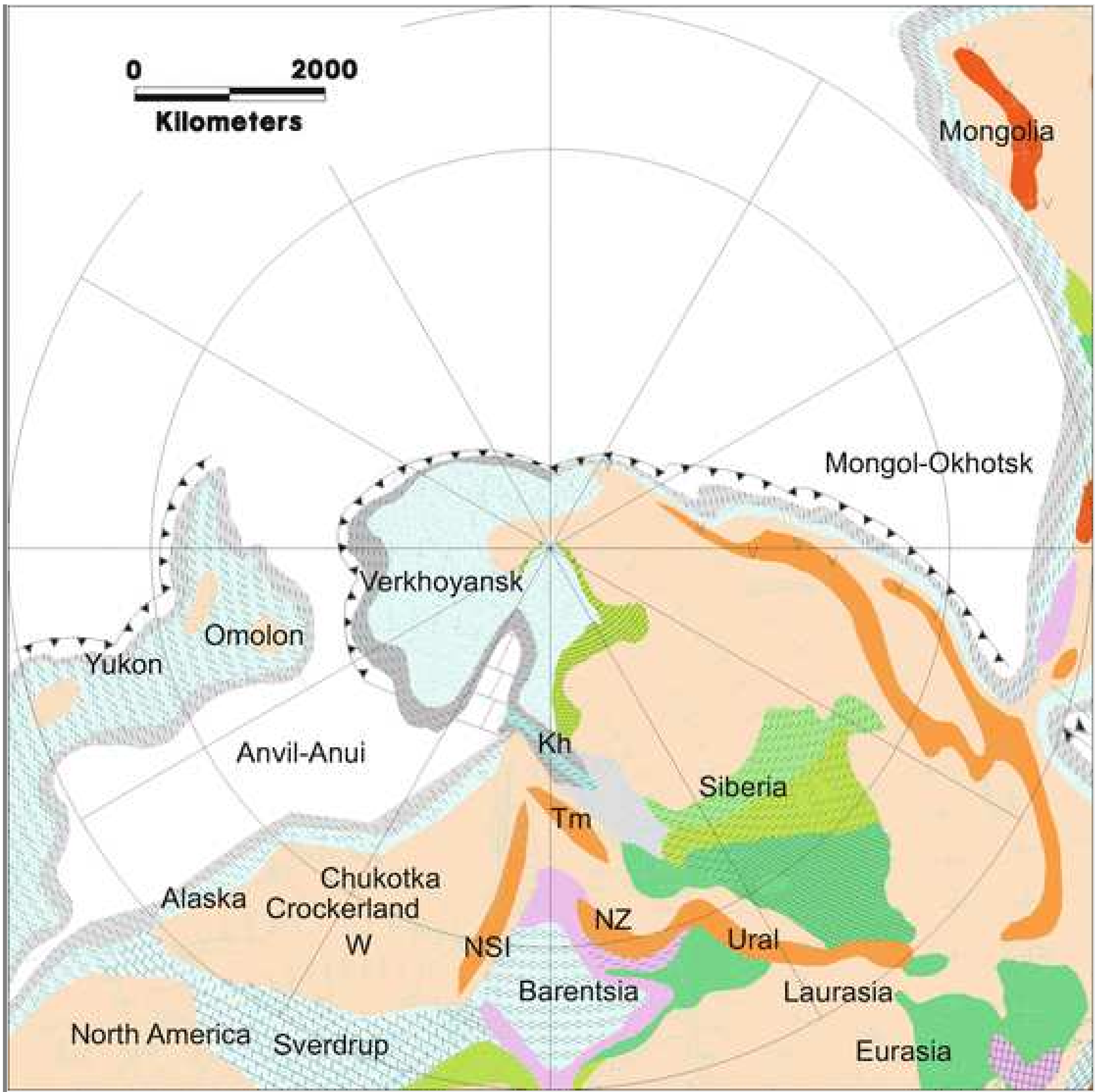
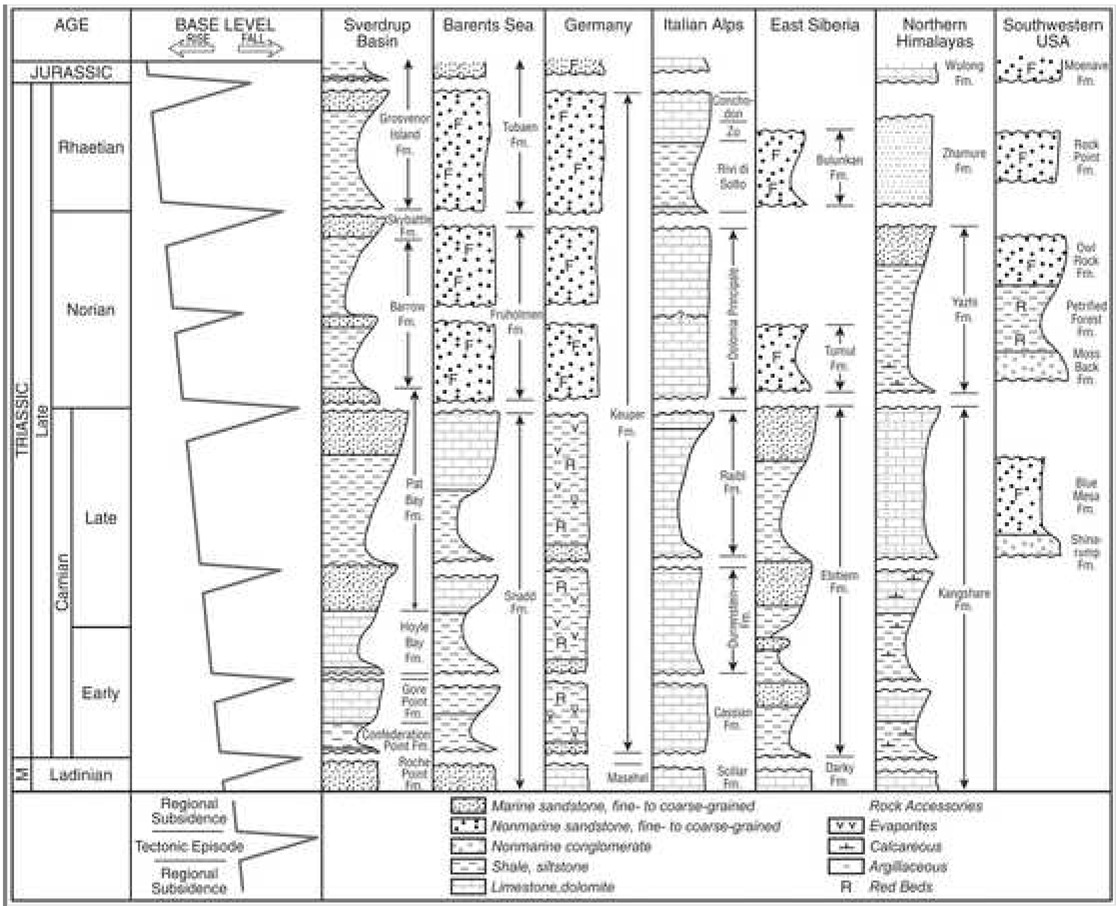


Fig. 2.9

Late Triassic stratigraphy of various basins contains sequence boundaries of basal Carnian, mid-Carnian, basal Norian, mid-Norian, basal Rhaetian, and latest Rhaetian age. The features of these boundaries indicate they represent relatively short-lived, tectonic episodes. Each tectonic episode was characterized by a rapid base level fall followed by rapid rise which punctuated the relatively slow, long term subsidence of the basins



AQ1
AQ2
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2.2. Methods

The Phanerozoic maps were constructed using a plate tectonic model that describes the relative motions between approximately 300 plates and terranes (Golonka 2000). This model was originally constructed using PLATES and PALEOMAP software, later the GPLATES program was used (see the detailed reconstruction methodology in Golonka et al. 2003b. The rotation file was presented in Golonka (2007a), and is shown in the appendix of that paper.

We modified this model using new paleomagnetic data, especially in the Tethys and Arctic areas (Kravchinsky et al. 2002; Hounslow and Nawrocki 2008; Kovalenko 2010; Metelkin et al. 2011, 2012; Uno et al. 2011; Domeier et al. 2012; Choulet et al. 2013; Vernikovsky et al. 2013; Wang et al. 2013; Song et al. 2015; Huang and Opdyke 2016; Li et al. 2016a, b; Zhou et al. 2016). We left the position of the major continent unchanged due to the absence of important new data. For example, according to Metelkin et al. (2011) there is an absence of authentic data for the Middle and Late Triassic from Siberia.

The facies were reconstructed using established sedimentological concepts for reefs and other sedimentary environments (Kiessling and Flügel 1999; Kiessling et al. 2003) and also presented by Golonka et al. (2006b) and Golonka (2007a, b). The calculated paleolatitudes and paleolongitudes were used to generate computer maps in Microstation design (.dgn format) converted later into Corel Draw (.cdr format). Facies and paleoenvironment information were posted after reviewing database files, regional paleogeographic maps and relevant papers. Information from several general and regional paleogeographic papers were filtered and utilized (Vinogradov 1968; Ziegler 1982, 1988; Hongzen 1985;

Ronov et al. 1989; Cook 1990; Zonenshain et al. 1990; Doré 1991; Dercourt et al. 1993, 2000; Golonka et al. 1994, 2006a; Metcalfe 1994, 2011, 2013a, b; Veevers 1994, 2006, 2013; Nikishin et al. 1996; Sengör and Natalin 1996; Puchkov 1997; Kiessling and Flügel 1999; Golonka 2000, 2002, 2007a, b, 2011; Golonka and Ford 2000; Ford and Golonka 2003; Scotese 2004; Miller et al. 2006; Robertson 2007; Feist-Burkhardt et al. 2008; Heydari 2008; Maurer et al. 2008; Miall and Blakey 2008; Miall et al. 2008; Pčelina and Korčinskaja 2008; Scheck-Wenderoth et al. 2008; Schmid et al. 2008; Peng et al. 2009; McKie and Williams 2009; Glørstad-Clark et al. 2010; Metelkin et al. 2011, 2012; Schettino and Turco 2011; Sibuet et al. 2012; Li and Huang 2013; Luo et al. 2014; Ershova et al. 2015a, b; Pease et al. 2015; Lane and Stephenson 2016; Müller et al. 2016; Toro et al. 2016; Cai et al. 2017; Centeno-García 2017).

2.3. Convergent Tectonics

The Late Paleozoic supercontinent Pangaea included North America, South America, Africa, Australia, Europe and Siberia and was surrounded by the Panthalassa Ocean (Figs. 2.1 and 2.2). The collision between Siberia and Europe formed the Ural Mountains during the Uralian Orogeny (Zonenshain et al. 1990; Nikishin et al. 1996; Puchkov 1997). The last episode of this orogeny occurred at the end of the Triassic in Novaya Zemlya (Toro et al. 2016; Zhang et al. 2017a). Deformation also affected the Taimyr Peninsula (Torsvik and Anderson 2002; Golonka 2007a, b). According to Zhang et al. (2017b) the Taimyr Permo-Triassic magmatic rocks were locally folded and faulted as a result of Late Triassic to Early Jurassic dextral transpression. According to Vernikovsky (1995) and Vernikovsky et al. (2003) the formation of the Taimyr structures is connected with the collision of the Kara microcontinent with Siberia. The uplift of the adjacent areas of Europe and Siberia was related to these orogenic events (Figs. 2.2 and 2.8). The Crockerland uplifted area of the Alaska-Chukotka micro-plate supplied sediments to the adjacent Sverdrup Basin in North America and was linked with Siberia at this time (Fig. 2.8; Anfinson et al. 2016). The subduction zones, known as the Late Paleozoic Pangaean Rim of Fire, were still active during the Triassic (Golonka and Ford 2000; Golonka 2002, 2004, 2007a, b; Matthews et al. 2016). This Rim of Fire was especially active along the western coast of Pangaea (Figs. 2.1 and 2.2). Active volcanism, terrane accretion, and back-arc basin development accompanied the subduction zones (Golonka 2007a, b). The subduction accompanied by magmatism was active in Central and North America (Goodge 1989, 1990; Dorsey and LaMaskin 2007; Centeno-García et al. 2008; Arvizu and Iriondo 2015) as well as in South America (Bustamante and Juliani 2011; del Rey et al. 2016). The movement of terranes within Panthalassa was related to the activity of this subduction (Figs. 2.1 and 2.2). According to Dorsey and LaMaskin (2007), the collision of terranes in North America happened during Late Triassic times in the Blue Mountains of Oregon. The position of these terranes is a subject of controversy, however (e.g. Engebretson et al. 1985; Panuska 1985; Debiche et al. 1987; Sengör and Natalin 1996; Keppie and Dostal 2001; Belasky et al. 2002; Trop et al. 2002; Piercey et al. 2006; Golonka 2007a, b; Colpron and Nelson 2011; Roniewicz 2013; Matthews et al. 2016). The relationship between Panthalassa terranes and Cimmerian plates was previously postulated and mapped (Golonka 2007a, b). The Panthalassa terranes bearing reef complexes were also mentioned by Flügel (2002). According to Peyberness et al. (2016, see also Stanley and Onoue 2015) the Western Panthalassa reefs from Japan corresponds with those of the Tethys Ocean during the Late Triassic. The Late Triassic was the time of the collisions now known as the Early Cimmerian and Indosinian orogenies. Blocks of the Cimmerian provenance and Eurasia (Sengör 1984; Sengör et al. 1984; Sengör and Natalin 1996) were involved in these collisions with the southern margin of Eurasia (Golonka 2000, 2002, 2007a, b; Golonka et al. 2003a, 2006a,b; Robertson 2007; Richards 2015). This series of collisions closed the Paleotethys Ocean. The closure happened earlier in the Alpine-Carpathian-Mediterranean area, later in the Eastern Europe-Central Asia and latest in the South-East Asia (Figs. 2.1 and 2.2). Microplates now included in the Alpine-Carpathian systems formed the marginal part of Europe. Subduction developed south of this zone. Late Triassic collisional events occurred also in the Moesia-Rhodopes areas (Tari et al. 1997; Golonka 2004, 2007a, b; Okay and Nikishin 2015; Petrík et al. 2016). The Alborz and the South Caspian

Microcontinent collided with the Scythian platform in Eastern Europe, and the other Iranian plates, including the large Lut block, collided with the Turan platform (Zonenshain et al. 1990; Kazmin 1991; Nikishin et al. 1996, 1998a; Golonka 2004, 2007a, b; Heydari 2008; Wilmsen et al. 2009; Masoodi et al. 2013; Okay and Nikishin 2015; Zanchi et al. 2009, 2016). Compressional deformations were recorded in the Caucasus, and Kopet Dagh areas, accompanied by the general uplift of the Fore-Caucasus, Caucasus and Middle Asia regions (Golonka 2004). According to Okay and Nikishin (2015), the accretion of an oceanic plateau was recorded by Late Triassic eclogites in the Pontides. Collisional events were also noted in Afghanistan and Pamir areas (Sengör 1984; Zonenshain et al. 1990; Golonka 2004, 2007a, b; Montenat 2009; Robinson 2015).

The Paleotethys between Qiangtang and Eurasia was closed during Late Triassic times (Figs. 2.1 and 2.2; Metcalfe 2013a; Zhai et al. 2013; Zhu et al. 2013; Luo et al. 2014; Song et al. 2015; Wu et al. 2016). The eastern Cimmerian plates were involved in the Indosinian orogeny. This name was derived from Indochina, the region where the orogeny was noted over one hundred years ago (Deprat 1913, 1914; Fromaget 1927, 1934, 1941, 1952). A major unconformity was observed in Northwest Vietnam. The deformed Lower – lowermost Upper Triassic (up to Carnian) marine metamorphosed rocks arranged into nappes and thrusts are covered by Upper Triassic continental red conglomerates ("terraines rouges", see Deprat 1913, 1914, also Golonka et al. 2006b). According to Lepvrier et al. (2004 see also Maluski et al. 2001, 2005; Lepvrier and Maluski 2008 and references therein), the main metamorphic event occurred during the Early Triassic, 250–240 Ma. The Late Triassic unconformity and 225–205 Ma postorogenic plutonism was noted by Faure et al. (2014). Hung (2010) describes magmatism in northeastern Vietnam related to Triassic Indosinian orogeny. According to Faure et al. (2014) the Jinshajiang and Ailaoshan belts in China and their geodynamic evolution, with Vietnam orogeny marking the same Indosinian Orogeny. It was related to the closure of Paleotethys Ocean along Raub-Bentong, Sra Kaeo and Nan-Uttaradit suture between Sibumasu and Indochina and Ailaoshan suture between Sibumasu and South China (Metcalfe 1994, 1996, 2000, 2011, 2013a, b; Golonka et al. 2006b and references therein).

One of the best examples of the Late Triassic orogenic event occurs in the Thailand/Myanmar trans-border zone. The Triassic-Jurassic succession in the Mae Sot area (northern Thailand), belongs to the Shan-Thai terrane. This block is subdivided into several zones from the west to east, including the Mae Sariang zone, where the Mae Sot area is located. This zone contains rocks of Triassic cherts (radiolarites), carbonates and flysch (turbiditic) facies, which indicate both pelagic condition and synorogenic deposits. From a paleogeographic point of view, the Shan-Thai block was a remnant of Paleotethys Ocean (Meesook and Sha 2010), which occupied a wide realm between Cimmerian Continent and Eurasian plate during Late Paleozoic-Early Mesozoic times. On the other hand, the Late Triassic Indosinian orogenic event has been associated with the docking and amalgamation of the Indoburma, Shan-Thai (Sibumasu) and Indochina terranes, which recently constituted the main part of Southeast Asia. Therefore, entire Jurassic units of these regions are represented by post-orogenic continental-shelf deposits, which are underlain discontinuously by older rocks. The oldest Jurassic bed, or the youngest Triassic bed, is the so-called "base-conglomerate", in local nomenclature, and is characterized by limestone and chert pebbles-bearing conglomerate, which is significant for the understanding of the tectonic evolution of the Shanthai terrane (Ishida et al. 2006; Meesook and Sha 2010). The underlying cherts are dated biostratigraphically (based on radiolarians) as Middle-Late Triassic. Limestone and chert pebbles from the "base-conglomerate" are dated as Early-Late Triassic by conodonts and as Middle-Late Triassic by radiolarians, respectively. These microfossils from pebbles constrain the age of the Indosinian (ShanThai = Mae Sariang) orogeny. Additionally, the youngest clasts, both limestones and siliceous rocks, indicate a strictly pelagic character of sedimentation up to Late Triassic time (see Ishida et al. 2006). A full open ocean condition must have existed at least before the end of the Triassic. The "base-conglomerate" is characterized by poorly-sorted, chaotically organized, pebble/fragment-bearing sedimentary breccia with no evidence of bivalve borings on their surfaces. The multicolored clasts are subrounded and subangular, and occur within reddish silt matrix. Chert clasts are red, green and grey and carbonate pebbles are represented both by micritic, pelagic limestones and the entire spectrum of packstones and grainstones, including extremely shallow-water bioclastic limestones (with bivalve fragments, crinoids, fragments of corals, etc.) with ooids and coated grains. The "base-conglomerate" is overlain by limestones and marls with mudstone intercalations of the Khun Huai Formation of the Hua Fai Group, dated by ammonites and bivalves as Early Toarcian. These facts indicate, by superposition, that the "base-conglomerate" is the latest Triassic or earliest Jurassic in age, according to the latest Triassic age of the chert and limestone pebbles within it. Sedimentological features indicate, on the other hand, a very rapid sedimentation event during its origin, such as erosion of steep, submarine "cliffs" that formed proximal aprons of debris flows. Additionally, the composition of this conglomerate, which has both deep-marine clasts and shallow-water ones, without any evidence of their long-distance transport, suggests erosion of different type of source material, which most probably originally took place in a different part of the primary Paleottethys Ocean. Then, they were removed, folded (forming nappes?) and overthrust to another location where they were destroyed and eroded, and produced marine molasse-type deposits unconformably overlying Indosinian deformed rocks. In fact, these data indicate both time and space reorganization of this orogenic system, which took place possibly during latest Triassic to earliest Jurassic time. The examination of the main orogenic events in the Southeast Asia regions indicates diachronous, multi-stages movements of the Indosinian orogeny. These include Early Triassic and Carnian/Norian orogenic pulses in Vietnam (Lepvrier et al. 2004), late Middle Triassic–early Late Triassic activity, the so-called second Indosinian event (Hahn 1984; Lepvrier and Maluski 2008, see also Cai et al. 2017) and close to the Triassic/Jurassic boundary in Thailand, as the Asian plate docked first on the East and later on the West (in modern coordinates).

Additionally, the Late Triassic volcanogenic-sedimentary event in Myanmar correlates presumably with synorogenic processes, which are represented by the Late Triassic flysch deposits with basaltic pillow lavas of the Shweminbon Group (Upper Triassic – Lower Jurassic turbidites), formerly part of Loi-an Group, the Bawgyo Group (Upper Triassic) and their equivalents, and with Upper Triassic turbidites represented by the Thanbaya/Pane Chaung Group/formations (Bannert et al. 2011; Win Swe 2012; Cai et al. 2017).

The collision between the South Chinese plate and the North Chinese block began during the Late Permian and continued during the Triassic (Yin and Nie 1996; Golonka et al. 2006b; Golonka 2007a, b). The Qinling orogenic belt records this collision. According to Dong et al. (2011) the Shangdan zone between the North and South Qinling belts is the suture separating the convergence and collision between North South Chinese plates. The post-suturing plutons were emplaced along the suture zone and on the adjacent plates (Bao et al. 2015; Liang et al. 2015; Lu et al. 2016). Consolidation of North China and Mongolia occurred mainly earlier but continued during the Triassic between North China and Mongolia. The newly formed, larger plate contains volcanics and collisional granites (Fig. 2.6; Chen et al. 2000; Wu et al. 2002; Shi et al. 2016). This consolidation left open a large embayment between Mongolia and Laurasia, the so-called Mongol-Okhotsk Ocean (Zonenshain et al. 1990; Golonka 2000, 2007a; Zeng et al. 2014). Active subduction existed along the margin of this ocean (Figs. 2.6 and 2.8), dipping cratonwards towards East Siberia (Zonenshain et al. 1990; Golonka 2007a, b), and granitic intrusions occurred along the Siberian margin (Zonenshain et al. 1990; Donskaya et al. 2013, 2016). The new, large Chinese-Southeast Asian plate including North and South China, Mongolia and eastern Cimmerian plates was consolidated at the Triassic-Jurassic Boundary (Fig. 2.2).

2.4. Extensional Tectonics

The onset of Pangaean break-up constitutes the main Late Triassic extensional tectonic event (Golonka 2007a, b). The rift basins originated between North America and Africa. The extensional rifting was accompanied by strike-slip faulting and block rotation (Ford and Golonka 2003; Laville et al. 2004;Golonka 2007a, b). Incipient continental rifting occurred also between northern Europe and North America (Fig. 2.2), reactivating the Late Paleozoic fracture system (Ziegler 1982; Doré 1991; Nikishin et al. 2002; Golonka 2011), and activating the North Sea rifts. The Central European Permian rift system known as the Polish/Danish Aulacogene was still active during Late Triassic times. The Upper Permian (Zechstein) salt went into salt tectonic phase with incipient salt diapirism and extrusion (Kutek 2001; Krzywiec 2012). Continental extension also began in isolated areas in South America during the Late Triassic (Macdonald et al. 2003; Ford and Golonka 2003; Golonka 2007a, b). Additionally, rift basins developed behind the subduction zone along the western Pangaean margin (Goodge 1989, 1990; Golonka and Ford 2000; Golonka 2007a, b; Centeno-García et al. 2008; Dickinson 2009; Bustamante and Juliani 2011; Giambiagi et al. 2011; Baby et al. 2013; Helbig et al. 2013; Spikings et al. 2016; Centeno-García 2017).

The Pangaea rift systems extended also to the Barents shelf, Arctic, and Siberia (Golonka 2011; Golonka et al. 2003a, 2006b). Rifting in Siberia was associated with the subduction zone at the Mongol-Okhotsk Ocean margin (Figs. 2.1, 2.2 and 2.8). Late Triassic sea-floor spreading in Siberia constituted an extension of the Anyui Ocean, which existed between the Alaska-Chukotka and Verkhoyansk terranes (Fig. 2.8; Zonenshain et al. 1990; Sengör and Natalin 1996; Golonka et al. 2003a; Golonka 2011). The opening of the Amerasia Basin appears to have begun near the Norian/Rhaetian boundary resulting in the rotational separation of the Alaska-Chukotka terrane from northern Laurasia (Embry and Anfinson 2014).

The volcanics (flows and intrusions) of the Central Atlantic Magmatic Province (CAMP), were emplaced at the end of Triassic and beginning of the Early Jurassic (e.g., Olsen 1997; Withjack et al. 1998; Marzoli et al. 1999, 2004, 2011; Knight et al. 2004; Golonka 2007a; Cirilli et al. 2009). CAMP constitutes one of the largest known Phanerozoic flood basalt provinces. It triggered climate changes and the end-Triassic extinction event (Wignall 2001; Lucas and Tanner 2008; Preto et al. 2010; Bond and Wignall 2014; Müller et al. 2016). The Late Triassic northward drift of the Cimmerian continent was accompanied by active seafloor spreading within the Neotethys Ocean. The spreading was driven by trench-pulling forces related to the north-dipping subduction, as well as the ridge-pushing forces related to mantle upwelling, expressed by hot spot activity (Golonka and Bocharova 2000; Golonka 2004, 2007a, b). Rifting and the opening of oceanic type basins could have occurred in the Alpine, Carpathian, Balkans and future Mediterranean area (Figs. 2.1, 2.2 and 2.3; Golonka et al. 2006a). The opening of the incipient Pindos–Maliac Ocean was related to the establishment of the Pelagonian, Sakariya and Kirsehir blocks as separate microplates within the Western (Robertson et al. 1991, 1996; Ferriere et al. 2016). The proto-Transylvanian and Vardar oceans originated within Carpathian-Balkan. The Tisa block was perhaps fully separated from the European margin by the Meliata-Halstatt Ocean. The positions of the Vardar, Meliata-Halstatt, Transylvanian, Pindos, Maliac oceans and their embayments within the Western Tethys remain quite speculative and are subjects of the debate (e.g., Kozur and Krahl 1987; Săndulescu 1988; Kozur 1991; Channell and Kozur 1997; Mock et al. 1998; Ivan 2002; Golonka 2004; Haas and Pero 2004; Golonka et al. 2006a; Dallmeyer et al. 2008; Schmid et al. 2008; Hoeck et al. 2009; Gawlick and Missoni 2015; Meinhold and Kostopoulos 2013). The Eurasian platform east of the Carpathians and Meliata Ocean was dissected by rifts that extended from the Dobrogea, through the proto-Black Sea area and along the margins of Scythianturan platform and probably were connected with Polish/Danish Aulacogene (Fig. 2.3; Zonenshain et al. 1990; Kazmin 1990, 1991; Nikishin et al. 1998a, b; Golonka 2004). The Tauric basin, which belonged to this rift system, was located between Pontides and the Dobrogea-Crimea segment of the Scythian platform (Golonka et al. 2006a). The North Dobrogea part of the rift zone separated Moesia and Eastern European platform (Muttoni et al. 2000; Golonka 2004; Golonka et al. 2006a). Several blocks were located between the rifted zone and the Neotethys (Golonka 2004; Golonka et al. 2006a; Okay and Nikishin 2015). This rifted zone can be interpreted as a back-arc basin resulting from the northward subduction of the Neotethys Ocean (Figs. 2.1, 2.2 and 2.3). The deep-water basin was located between Apulia, the Taurus platform and the African continent (Fig. 2.1; Catalano et al. 1991; Kozur 1991; Marsella et al. 1993; Golonka 2004, Golonka et al. 2006b). It was connected eastwards with an oceanic-type basin recorded by the Mamonia ophiolites complex in Cyprus (Robertson and Woodcock 1979; Morris 1996; Robertson 1998). The rifts cutting Apulia were connected with the western part of Neotethys.

The whole Paleotethys was closed in the western part of the Tethyan realm in the Early Jurassic (Fig. 2.2). The pulling force of the north-dipping subduction along the northern margin of Neotethys caused the drift of a new set of plates from the passive Gondwanian margin. These plates divided the Neotethys Ocean into northern and southern branches (Golonka 2004). Metcalfe (2013a) distinguished Cenotethys as the southern branch. The Lhasa block was the most prominent plate which drifted away from Gondwana (Sengör 1984; Dercourt et al. 1993; Metcalfe 1994, Metcalfe 2013a, b; Sengör and Natalin 1996; Yin and Nie 1996; Golonka 2004; Cai et al 2016; Li et al. 2016a, b; Lu et al. 2016; Meng et al. 2016; Zhou et al. 2016). According to Li et al. (Li et al. 2016a, b), the Kirsehir, Sakarya (Robertson et al. 1991, 1996), and perhaps the Lesser Caucasus-Sanandaj-Sirjan, Biston-Avoraman plates drifted in the central Neotethys area (Adamia 1991; Robertson et al. 1991, 1996, 2004; Arfania and Shahriari 2009; Mehdipour Ghazi and Moazzen 2015; Nouri et al. 2016). According to Metcalfe (2013a, b) South West Borneo and East Java-West Sulawesi were separated from Northwest Australia in the Late Triassic in the easternmost Tethys area. The consolidation of the Chinese and southeastern Asian blocks was followed by extensional tectonics caused by the pulling force if the new Neotethys subduction. Consequently, rift basins developed in China and adjacent areas (Golonka et al. 2006b; Luo et al. 2014). This process was enhanced by the Panthalassa (Paleo-Pacific) plate sliding beneath the Eurasian plate (Luo et al. 2014; Li et al. 2016a, b).

2.5. Sedimentation and Paleolithofacies

Continental rifts, which originated during Triassic times, were filled with clastic deposits, particularly abundant red beds consisting of fluvial deposits and accompanied by evaporites (Ziegler 1988; Withjack et al. 1998; Golonka and Ford 2000; Kutek 2001; Feist-Burkhardt et al. 2008). Mixed siliciclastics, carbonates and evaporates were deposited in Central Europe (Figs. 2.3 and 2.9) as to the upper part (Keuper) of the Central European tripartite facies sequence that gave the Triassic its name (Köppen and Carter 2000). The Keuper Formation encompasses the Carnian, Norian and Rhaetian stages (Fig. 2.9). The accumulation of sediments in this area reached up to 4000 m due to significant subsidence (Köppen and Carter 2000; Kutek 2001; Golonka 2007a, b; Feist-Burkhardt et al. 2008). Meanwhile, continental red beds were deposited in the eastern United States while a marine shelf existed on the western North America margin. Continental rifting occurred between northern Europe and Greenland (Fig. 2.8). The Pangaea rift systems extended from the Newark and Central Europe basins through the North Atlantic, to the Barents shelf and Arctic Alaska (Figs. 2.3, 2.8 and 2.9). These rifts were filled primarily with red continental clastics reflecting arid climate (Ronov et al. 1989; Olsen 1997; Golonka et al. 2003a, 2006a, b; Golonka 2007a, b; Dickinson 2004, 2008, 2009; Miall et al. 2008; Miall and Blakey 2008). Carbonate sedimentation dominated in the Alps and Carpathians (Golonka 2004, 2007a, b; Feist-Burkhardt et al. 2008). This sedimentation was associated with existence of platforms on the Neotethys and Paleotethys margins as well as on Cimmerian microplates. Shallow-water limestones and dolomites with algal/coral-dominated reefs were deposited on these platforms (Golonka 2007a, b). They were accompanied by fine grained clastics (Figs. 2.3 and 2.9). Many of the western Tethyan reefs were located on these platforms. Triassic carbonate platforms and reefs were formed not only in the Tethys, but also in the western and eastern parts of the Panthalassa (Paleo-Pacific) Ocean (Golonka 2007a, b). A large carbonate platform that spread from Apulia to the Taurus zone provides an example (Dercourt et al. 1993, 2000; Golonka 2004, 2007a, b; Feist-Burkhardt et al. 2008) in that it contains significant numbers of reefs (Kiessling and Flügel 1999; Flügel 2002) and was connected with the Alpine-Inner Carpathian carbonate platforms, which also contained abundant reefs (Kiessling and Flügel 1999; Flügel 2002). Dolomitization of the platform limestones was common and dolomites are widespread in Southern Europe and Central Asia. The Dolomia Principale (Fig. 2.9) represents a classic example of the Tethyan dolomites. The Dolomites range in the Italian Southern Alps took their name from the mineral and rock dolomite, which in turn were named after the French geologist Dieudonné Sylvain Guy Tancrède de Gratet de Dolomieu by de Saussure (1792). Dolomites were also widespread on the southern margin of Eurasia in the Caspian area and in Central Asia (Figs. 2.5 and 2.7). Continental and marginal marine sediments with evaporites and volcanics were also deposited in this part of Eurasia (Zonenshain et al. 1990; Dercourt et al. 1993, 2000; Nikishin et al. 1996, 1998a, b; Brunet et al. 2002; Zharkov and Chumakov 2001; Golonka 2004, 2007a, b). The neritic and lagoonal sediments of so-called Carpathian Keuper were deposited in the Northern Carpathians during the latest Triassic, marking the uplift of the Inner Carpathian plate (Kotański 1961; Golonka 2004; Feist-Burkhardt et al. 2008; Rychliński 2008). The Neotethyian margins of Greater India, Arabia and Australia (Figs. 2.5 and 2.7) were occupied by mixed carbonate-clastic facies (Cook 1990; Alsharhan and Magara 1994; Golonka and Ford 2000; Golonka 2007a, b). Basins containing Triassic continental red bed deposits were located in Gondwana (Fig. 2.7), in South America, Africa, Antarctica, Madagascar and India (Golonka 2007a, b). The deposition of synorogenic flysch sequences in South-East Asia was linked to the Indosinian orogenic collisional events (Hahn 1984; Golonka et al. 2006b; Lepvrier and Maluski 2008; Cai et al. 2017). They were accompanied by pelagic cherts, cherty limestones and fine-grained clastics as well as by volcanoclastics and pillow lavas (Ishida et al. 2006; Bannert et al. 2011; Win Swe 2012; Cai et al. 2017) and followed by post-orogenic molasses. Post-orogenic Upper Triassic continental red conglomerates are known as “terraines rouges” of Deprat (1913, 1914) in Vietnam (Golonka et al. 2006a, b). These red-bed postorogenic facies that follow synorogenic turbidites and pelagic cherts are also known from the Malaysian Peninsula (Oliver and Prave 2013; Ridd 2013). Flysch sequences and volcanoclastic deposits occur on the Lhasa plate, South Tibet (Liu et al. 2012). Shallow marine, carbonate and clastic sedimentation dominated on the Qiantang plate (Zhu et al. 2013; Wu et al. 2016). Various stratigraphic sequences representing different paleogeographic facies existed in South China where paralic clastics, shallow marine clastics, shelf carbonate platform facies and deep water turbidites can be distinguished. Siliciclastic sedimentation prevailed in North China, including shallow marine clastics, marginal marine deposits such as deltas, as well as turbidites accompanied by volcanoclastics (Fig. 2.6; Hongzen 1985; Tong and Yin 2002; Golonka 2007a, b; Cao et al. 2010; Luo et al. 2014; Li et al. 2014). The sediments consisting mainly of fine-grained molasse-type filled the foreland basins following the Uralian orogeny in Timan-Pechora, Novaya Zemlya and eastern Barents regions. Siliciclastics were common in the Siberia and Arctic regions (Figs. 2.8 and 2.9; Embry 1988, 1993, 1997; Nikishin et al. 1996; Golonka and Ford 2000; Golonka et al. 2003a; Golonka 2007a, b, 2011; Toro et al. 2016); the Sverdrup Basin of Arctic Canada was a main depocenter with the Late Triassic succession of fluvial to marine slope deposits being over 2500 m thick (Embry 1997). Triassic, restricted-marine shelf basins contain black shales that have source rock potential (Leith et al. 1993; Golonka et al. 2003a; Golonka 2007b). Upper Triassic source rocks, important for hydrocarbon exploration in the North Atlantic, were identified in the Jameson Land Basin, East Greenland (Andrews et al. 2014).

2.6. Global Base-Level Changes

In this section, we briefly review postulated base-level changes that have been interpreted to affected numerous basins throughout Pangaea during the Late Triassic. We first look at small scale changes with frequencies of less than 500,000 years. Then we address large scale base-level changes with frequencies of greater than 2 million years.

Tanner (2010) comprehensively reviewed the literature for high frequency, small scale cycles for the entire Triassic. Such cycles have been recorded in various Late Triassic successions with the best documentation being from the rift valley deposits of the Newark Group of the northeastern USA (Olsen and Kent 1996) and the carbonate platforms of the Italian Alps (Cozzi et al. 2005; Schwarzacher 2006). The Late Triassic small scale cycles of the Italian Alps are characterized by the presence of exposure surfaces and paleosols. This leaves little doubt as to such cycles being generated by base-level changes caused by either eustasy or tectonics. It must be noted that any high frequency, small scale cycles which do not include exposure surfaces may well have an auto-cyclic explanation for their generation.

Given the occurrence of such high-frequency base level changes in the Late Triassic of the Italian Alps and the apparent coincidence of the calculated frequencies with those of the Milankovitch spectrum (Cozzi et al. 2005; Schwarzacher 2006), it seems reasonable to assume that small-scale, global sea-level changes driven by climate changes characterize the Late Triassic. However, as cautioned by Tanner (2010), this interpretation cannot be considered as unassailable for two main reasons. Given the greenhouse climate of the Late Triassic (Preto et al. 2010) and the consequent unlikelihood that substantial amounts of water could have been stored as ice during cold periods, there are no obvious mechanisms for climate changes to drive eustatic sea level change of the magnitude seemingly recorded by the cycles. The other problem is the general lack of precise radiometric age dates to constrain the interpreted cycle periods.

More studies are needed for Late Triassic, very shallow water carbonate and siliciclastic strata in a number of basins of Pangaea to see if they are characterized by high-frequency cycles that are capped by exposure surfaces. If Milankovitch climate change cycles were operating during the Late Triassic, then such cycles should be present in the successions of most, if not, every basin. In summary, it is quite possible that Milankovitch climate cycles were operating during the Late Triassic but further studies are needed to confirm or deny such a phenomenon.

Large scale, base-level changes are recorded in most Late Triassic successions and are expressed as large-magnitude, sequence boundaries. Such boundaries are characterized by an extensive unconformable portion on the basin margins and are the product of base-level changes that can exceed 100 m. Both eustatic and tectonic explanations have been offered for the generation of these boundaries.

Late Triassic, large-magnitude, sequence boundaries, which have been recorded in different basins throughout Pangaea, have been biostratigraphically dated as near the base Carnian, mid-Carnian, near the base Norian, mid-Norian, near the base Rhaetian and latest Rhaetian. Initially, these boundaries were interpreted to be the product of eustasy, including a significant sea level fall followed by sea level rise (Haq et al. 1987, 1988; Embry 1988; De Zanche et al. 1993; Gianolla and Jacquin 1998). Given a climate change/continental glaciation explanation was not possible, the authors appealed to changes in the volume of the world ocean (tectono-eustasy) as the main driver of such large scale eustatic changes.

Embry (1989, 1997) reversed his earlier interpretation and postulated that the large-magnitude sequence boundaries, which punctuated the entire Mesozoic succession of the Sverdrup Basin of Arctic Canada, were of tectonic origin. This interpretation was based on various characteristics of such boundaries which strongly favor a tectonic origin. Such characteristics included:

- A widespread, often angular, unconformity on the basin margins and positive elements
- A major change in depositional regime
- A notable change in tectonic regime and subsidence pattern
- A change in provenance for siliciclastic sediments
- A widespread transgression with significant deepening directly following the boundary.

Furthermore Embry (1997) demonstrated that the five, Late Triassic large-magnitude sequence boundaries present in the Sverdrup Basin are also present in basins in western Canada, southwestern USA, Barents Sea, Germany, Italian Alps, western Siberia, and northern Himalayas (Fig. 2.9). Notably, the unconformities in all these areas exhibit characteristics which favor a tectonic origin.

To explain the occurrence of simultaneous tectonic episodes in multiple and widely separated basins of Pangaea, Embry (1997) invoked the tectonic model of Cloetingh et al. (1985). The widespread, large magnitude base-level changes of the Late Triassic were interpreted to be an expression of relatively rapid and substantial changes in the horizontal and vertical stress fields that affected the Pangaea supercontinent (Fig. 2.9). Such stress changes would be possibly due to somewhat abrupt changes in the speed and/or direction of the plate movements that episodically affected Pangaea. Notably, it is possible that secondary tectono-eustatic effects were associated with such plate tectonic reorganizations (Embry 1997).

2.7. Climate Change and Episodic Tectonism

The climate of the Triassic has been reviewed by Preto et al. (2010) and they have interpreted that it "was characterized by a non-zonal pattern, dictated by a strong global monsoon system with effects that are most evident in the Tethys realm". For the Late Triassic, Preto et al. (2010) postulated that the monsoonal climate had its maximum expression and that there were three climatic zones which did not have a clear latitudinal distribution. These three zones included a dry climate for the western margin of Tethys and the central part of Pangaea, a wet and dry climate for the coasts of eastern Laurasia and Gondwana and the western coasts of Pangaea, and a wet climate in the high latitudes.

Although, in general, there was not much variability in climate throughout the Late Triassic, significant climate changes seem to be associated with the five tectonic episodes discussed in the last section. The most well-known of these is the "Carnian Pluvial Episode" (Ruffell et al. 2015) which corresponds with the mid-Carnian tectonic episode. This event was marked by warmer, more humid conditions in various parts of Pangaea and a notable increase of siliciclastic supply to numerous basins (Ruffell et al. 2015). Climate changes seem to have occurred associated with the other four tectonic episodes as shown by the marked changes in spore/pollen ratios associated with these boundaries (Hochuli and Vigran 2010). Climate change associated with the latest Rhaetian has been documented by various workers as summarized by Preto et al. (2010). The CAMP flood basalts, which were associated with the extensional phase of the latest Rhaetian tectonic episode, produced enormous amount of $CO_{2,}$ triggered global warming, and increased ocean acidification. These factors caused the end of Triassic extinction event

(Wignall 2001; Lucas and Tanner 2008; Preto et al. 2010; Bond and Wignall 2014; Müller et al. 2016).

2.8. Concluding Summary

Herein, we present a new set of global and regional paleogeographic maps for the Late Triassic (Carnian-Rhaetian) time interval. The global maps depict the plate tectonic configuration, present day coastlines, subduction zones, selected transform faults, spreading centers and rifts during the beginning (224 Ma) and end (200 Ma) of Late Triassic. The regional maps illustrate the Late Triassic paleoenvironment and paleolithofacies distribution for most important regions. The stratigraphic chart shows Late Triassic stratigraphy of various basins and sequence boundaries of basal Carnian, mid-Carnian, basal Norian, mid-Norian, basal Rhaetian, and latest Rhaetian age.

The Late Triassic was a time of collisional events, now known as Early Cimmerian and Indosinian orogenies. This series of collisions closed the Paleotethys Ocean. The closure happened earlier in the Alpine-Carpathian-Mediterranean area, later in the Eastern Europe-Central Asia and latest in the South-East Asia. The Indochina, Southeastern Asian and Qiangtang plates were sutured to South China. The new, large Chinese-Southeast Asian plate, including the North and South China, Mongolia and eastern Cimmerian plates, was consolidated at the Triassic-Jurassic Boundary. This consolidation left open a large embayment of Panthalassa, between Mongolia and Laurasia, known as Mongol-Okhotsk Ocean. The Uralian Orogeny, which sutured Siberia and Europe continued during Late Triassic times and was recorded in Novaya Zemlya.

The onset of the break-up of Pangaea constitutes the main Late Triassic extensional tectonics event. Continental rifts, which originated during this event, were filled with clastic deposits. Abundant red beds, accompanied by fluvial deposits and evaporites, were deposited in classic sedimentary systems. The pulling force of the north-dipping subduction along the northern margin of Neotethys caused the drift of a new set of plates from the passive Gondwana margin. These plates divided the Neotethys Ocean. Carbonate sedimentation was associated with existence of platforms on the Neotethys and Paleotethys margins as well as on Cimmerian microplates. Synorogenic turbidites and postorogenic molasses were associated with the Indosinian orogeny. The late stages of the Uralian orogeny in Timan-Pechora, Novaya Zemlya and eastern Barents regions included the filling of the foreland basin with fine-grained, molasse sediments. Siliciclastics were common in the Siberia and Arctic regions.

The widespread, large magnitude, base level changes of the Late Triassic are interpreted to be an expression of relatively rapid and substantial changes in the horizontal and vertical stress fields that affected the Pangaea supercontinent. Such stress changes would be possibly due to somewhat abrupt changes in the speed and/or direction of plate movements, which episodically affected Pangaea. The Late Triassic climate changes seem to be associated with the main tectonic episodes. The most well-known of these is the “Carnian Pluvial Episode” which corresponds with the mid-Carnian tectonic episode. The Central Atlantic Magmatic Province flood basalts, which were associated with the extensional phase of the latest Rhaetian tectonic episode, produced enormous amounts of CO_2, triggering global warming, increasing ocean acidification, and causing the latest Triassic extinction event.

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