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PHANEROZOIC PALEOENVIRONMENT AND PALEOLITHOFACIES MAPS. CENOZOIC

Mapy paleośrodowiska i paleolitofacji fanerozoiku. Kenozoik

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Abstracts: The paper presents the detailed plate tectonic, paleogeographic, paleoenvironment and plaeolithofacies maps for seven Cenozoic time intervals. Thirty five maps, generated using PLATES and PALEOMAP programs, contain information about plate tectonics, paleoenvironment, and paleolithofacies during Paleocene, Eocene, Oligocene, Miocene and Pliocene, time slices. The spatial reconstruction of basin architectures during their origin, expansion, closure and inversion as well as the dynamic of intrabasinal ridges were obtained by palinspastic modeling. This modeling utilized paleomagnetic data and stratigraphic-facies analysis of basins and ridges. Information contained within global and regional papers were selected and posted on the maps. The detailed paleoenvironment and plaeolithofacies zones were distinguished within the basins. The paleogeographic maps illustrate the geodynamic evolution of Earth from Late Cretaceous to Neogene, spreading and origin of new oceans, oceans closing, collisions, continents accretion and origin of new supercontinents.

Keywords: Paleocene, Eocene, Oligocene, Miocene, Pliocene, plate tectonics, paleogeography

Treść: Artykuł przedstawia szczegółowe mapy paleogeograficzne dla siedmiu przedziałów czasowych w obrębie kenozoiku. Trzydzieści pięć map, skonstruowanych przy użyciu programów PLATES i PALEOMAP, zawiera informacje dotyczące tektoniki płyt, paleośrodowiska i paleolitofacji w czasie paleocenu, eocenu, oligocenu, miocenu i pliocenu. Przestrzenną rekonstrukcję architektury basenów w okresie ich powstawania, ekspansji, zamykania i inwersji oraz analizę dynamiki grzbietów śródbasenowych uzyskano, wykonując modelowanie palinspastyczne, przy uwzględnieniu badań paleomagnetycznych oraz analizy stratygraficzno-facjalnej basenów i rozdzielających je grzbietów. Informacje zawarte w szeregu globalnych i regionalnych prac zostały wyselekcjonowane i naniesione na mapy. W obrębie basenów wydzielono poszczególne strefy paleośrodowiskowe i paleolitofacjalne. Mapy paleogeograficzne ilustrują geodynamiczną ewolucję Ziemi od późnej kredy po neogen, rozrost (spreding) i tworzenie się oceanów, zamykanie się oceanów, kolizje, łączenie się kontynentów i tworzenie się nowych superkontynentów.

Słowa kluczowe: paleocen, eocen, oligocen, miocen, pliocen, tektonika płyt litosfery, paleogeografia

INTRODUCTION

The aim of this paper is the presentation of Cenozoic paleographic maps of the world, containing paleoenvironment and paleolithofacies details. In the previous papers (Golonka 2007b, c, d), the author presented global paleogeographic maps for Late Paleozoic and Mesozoic. Now, the author attempts to cover the entire Phanerozoic in four papers. This paper is dealing with Paleocene, Eocene, Oligocene, Miocene and Pliocene time slices. The maps were constructed using a plate tectonic model, which describes the relative motions between approximately 300 plates and terranes. The detailed reconstruction methodology was described previously in Golonka *et al*. (2003a). The rotation file was presented in Golonka (2007b) paper, online version, the appendix. The facies were assembled according to rules established during the production of Phanerozoic reefs map (Kiessling *et al*. 1999, 2003) and also presented by Golonka *et al*. (2006) and Golonka (2007b, c, d).

MAP DISCUSSION

Late Cretaceous-Early Paleocene

The lithofacies and paleoenvironment data for Late Cretaceous-Early Paleocene maps (Figs 1–5) cover rather a large time span 81–58 Ma (Golonka & Kiessling, 2002) from the lower Campanian unconformity to the lower Thanetian unconformity. The Cretaceous--Paleogene boundary is contained within this supersequence, which began with a high sea--level, which slowly lowered, then dropped dramatically at the Danian-Thanetian boundary, while the global significance of the supposed regression at the Cretaceous-Tertiary boundary is a subject of controversy (Kiessling & Claeys 2000, Golonka & Kiessling 2002). For example, the detailed study of the Carpathian realm revealed formations and megasequences covering Maastrichtian and Early Paleocene without visible Cretaceous-Paleogene boundary (Golonka & Picha 2006, Golonka & Waśkowska-Oliwa 2007, Rasser *et al*. 2008).

The uplifted African Atlantic margin created internal drainage and narrow continental margins. The lower Thanetian unconformity is related to the inversion in Europe and convergence in Tethys (Golonka & Kiessling 2002). The Late Cretaceous-Early Paleocene time slice has been mapped on the 76 Ma plate tectonic reconstruction. It is equivalent to the Upper Zuni IV slice (Golonka 2000, 2002) upper part of Zuni III sensu Sloss (1988) and UZA-4 to TA1 of Haq *et al*. (1988).

During the latest Cretaceous, the spreading of the Central and Southern Atlantic continued (Figs 1, 2), with a significant increase in the size of the equatorial Atlantic (Nürnberg & Müller 1991, Ford & Golonka 2003, Golonka 2007d). The Atlantic passive margins were partially uplifted (Wernicke & Tilke 1989). Relatively narrow continental margins were established along Central Atlantic and equatorial Atlantic. Clastics prevailed along the equatorial Atlantic margins, locally, on the South American site with carbonates and with organic-rich rocks (Ronov *et al*. 1989, Ford & Golonka 2003, Kiessling *et al*. 2003, Summa *et al*. 2003) (Fig. 2 on the interleaf)). The similar patterns area also visible within passive margins along Africa (Bensalem 2002).

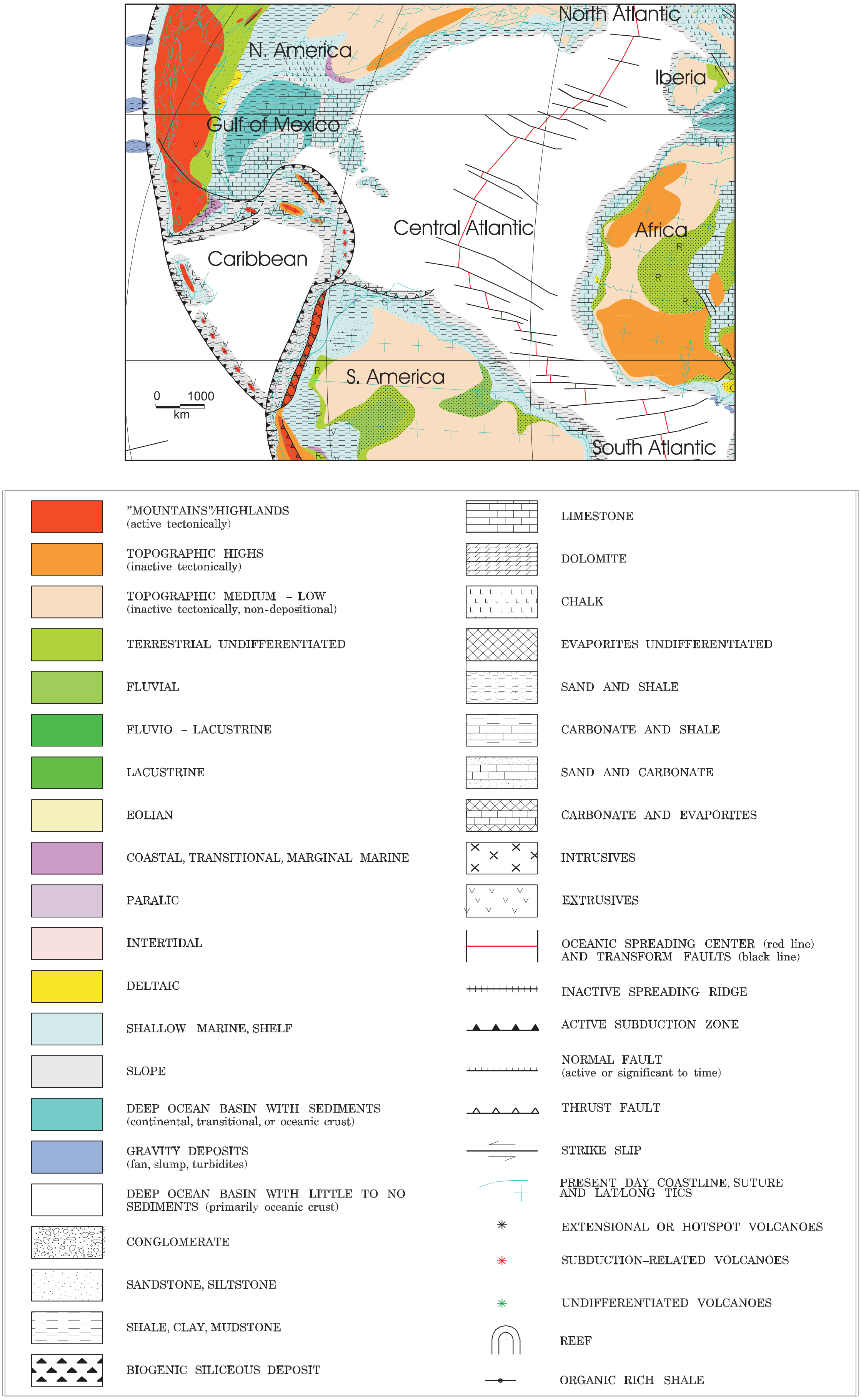

Fig. 2. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Late Cretaceous-Early Paleocene times. Explanations to figures 2–5, 7–10, 12–15, 17–20, 22–25, 27–30, 32–35. Qualifiers: B – bauxites/laterites, C – coals, E – evaporites, F – flysh, Fe – iron, G – glauconite, M – marls, O – oolites, P – phosphates, R – red beds, Si – silica, T – tillites, V – volcanics

Fig. 2. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich w późnej kredzie-wczesnym paleocenie. Objaśnienia do figur 2–5, 7–10, 12–15, 17–20, 22–25, 27–30, 32–35. Oznaczenia literowe: B – boksyty/lateryty, C – węgle, E – ewaporyty, F – flisz, Fe – żelazo, G – glaukonit, M – margle, O – oolity, P – fosfaty, R – utwory czerwone, Si – krzemionka, T – tillity, V – utwory wulkaniczne

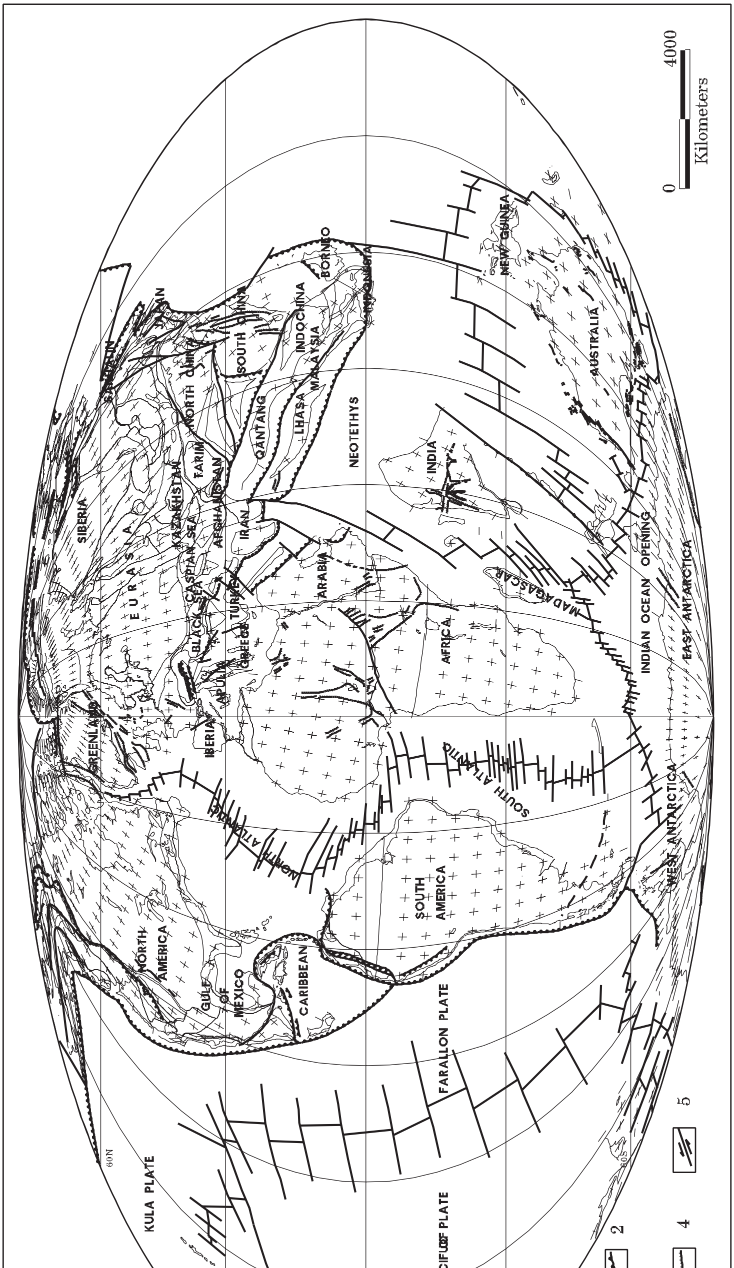

Fig. 1. Plate tectonic map of the Late Cretaceous-Early Paleocene (plates position as of 65 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 1. Mapa tektoniki płyt późnej kredy-wczesnego paleocenu (pozycja płyt 65 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

The Niger delta supplied large quantities of clastics (Ford & Golonka, 2003, Deng *et al*. 2008, Olowokere 2008). Carbonates developed on the margins of Central Atlantic in Africa and in North America in South Carolina, Florida, Bahamas and Yucatan (Ronov *et al*. 1989, Ford & Golonka 2003, Kiessling *et al*. 2003, Wilson 2008). Deep-water carbonates developed within the Gulf of Mexico. Chalks were present in the North America seaway north of Gulf of Mexico (Sloss *et al*. 1960, Ford & Golonka 2003).

The Caribbean arc was moving eastward toward the Antilles island arc causing subduction of the Proto-Caribbean crust (Ross & Scotese 1988, Golonka 2000, 2002, Golonka *et al*. 2006, Iturralde-Vinent 2006, García-Gasco 2008). This arc collided with the Bahama platform during the latest Cretaceous (Golonka 2007d). This collision caused the capture of the Caribbean plate, which was formed from the trapped part of the Farallon plate of the Pacific (Lawver & Gahagan 1993, Golonka 2007d). Subduction along the Panama Arc was also initiated (Scotese 1991). Cuba was in front of the Caribbean arc followed by Hispaniola. Large seaway existed between South America and North America (Figs 1, 2). The Amiache--Chaucha terrane was accreted to South America (Pindell & Tabutt 1995, Cediel *et al*. 2005). The western North American Cordillera continued to compress during the latest Cretaceous until the Eocene. This compression resulted in thrusting and margin-parallel, transcurrent faulting (Oldow *et al*. 1990). Large transform fault existed south of Yucatan, cutting across Mexican terranes (Golonka 2000, 2002, Ford & Golonka 2003, Keppie & Morán-Zenteno 2005, Ortega-Gutíerrez *et al*. 2007). Clastic sedimentation prevailed in the Caribbean area and along the western margins of North America with a significant share of volcanics and volcanoclastics. Carbonates developed locally on Cuba (Fig. 2).

Arabian plate was moving together with Africa northwards closing between its margin, Sanandaj-Sirjan and Lut plates (Robertson & Searle 1990, Ricou 1996, Golonka 2000, 2002, Giraud *et al*. 2005, Golonka *et al*. 2006, Pasyanos & Nyblade 2007, Allen & Armstrong 2008, Fakhari *et al*. 2008, Heydari 2008) (Figs 1, 3). The plate convergence was associated with obduction of ophiolites on the Arabian margin. The exotic rocks of the Oman ophiolitic nappes reflect different stages of the evolution of the Tethyan Ocean and its branches from the Permian to the Cretaceous (Pillevuit *et al*. 1997). Carbonate platforms existed on Arabia, adjacent part of Somalia and on the Lut plate (Ronov *et al*. 1989, Dercourt *et al*. 1993, Guiraud & Bellion 1996, Kiessling *et al*. 2003, Golonka 2004, Guiraud *et al*. 2006, Wilson 2008, Scheibner & Speijer 2008).

The Indian plate was located between latitude 30 degrees South and equator (Royer & Sandwell, 1989, Lawver *et al*. 1992, Golonka 2000, 2002, Kumar *et al*. 2007, Ali & Aitchison 2008) (Figs 1, 3). Its northward movement followed opening of the Mascarene Basin and separation with Madagascar (Scotese 1991, Golonka 2000, 2002), wide open Indian Ocean and narrowed significantly the Neotethys in this region. The emplacement of large volumes of flood basalts on India (Fig. 3) was associated with the movement of this plate over the Réunion hot spot (Baksi 1987, Coffin & Eldholm 1994, Golonka & Bocharova 2002, Mahoney *et al*. 2002, Kumar *et al*. 2007, Keller *et al*. 2008, Sen *et al*. 2008). Rifts caused by the mantle activity originated within the Indian plate. Carbonates and mixed carbonates-clastics prevailed on the plate's margins and on this part of Greater India now involved in the Himalayan structure (Kiessling *et al*. 2003, Acharyya 2007, Green *et al*. 2008). At the same time clastics, mainly accretionary prism flysch deposits were abundant along the northern margin of Neotethys (Liu & Einsele 1994, Wan *et al*. 2002, Zhang *et al*. 2002, Golonka *et al*. 2006). Uplift condition continued within South China, Indochina and Indonesia plates with continental deposits, red beds evaporites and volcanics (Golonka *et al*. 2006) (Fig. 3). North and west-dipping subduction rimmed Indonesia, while south-dipping subduction developed along the southern margin of the proto--South China Sea in South-East Asia north of Borneo (Fig. 3) (Lee & Lawver 1994, Golonka *et al*. 2006). Volcanics were abundant in South China. According to Chung *et al*. (1997), the Paleogene volcanic activities resulted from the lithospheric extension in South China that migrated southwards and eventually led to the opening of the South China Sea during later time.

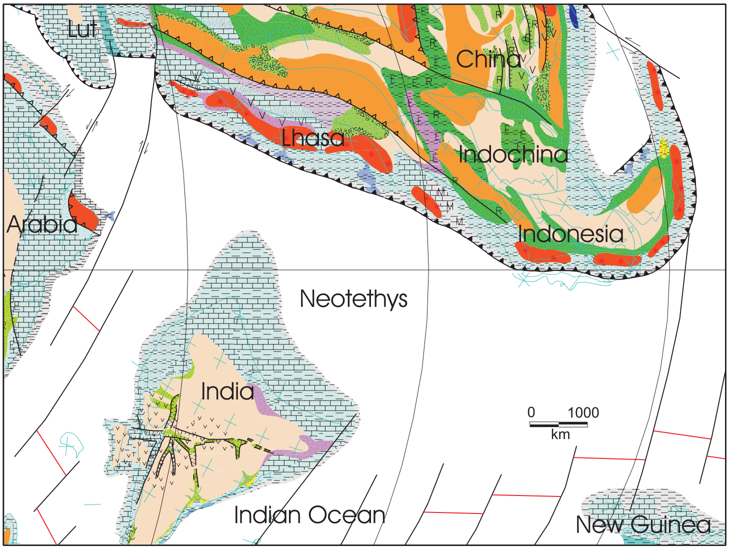

Fig. 3. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, eastern Neotethys and adjacent areas during Late Cretaceous-Early Paleocene times

Fig. 3. Mapa tektoniki płyt, paleośrodowiska, i litofacji Azji południowo-wschodniej, wschodniej Tetydy oraz obszarów sąsiednich w późnej kredzie-wczesnym paleocenie

Most of Western Europe was covered by a shallow sea (Fig. 4). Chalk was still widespread in western, central and southeastern Europe during the latest Cretaceous (Ziegler 1982, 1988, 1990). Large areas of Central Europe were uplifted due to Subhercynian and Laramide tectonic activities (Ziegler 1988, 1990, 1992, Baldschuhn *et al*. 1991, Mazur *et al*. 2005, Krzywiec 2006). This uplift was related to the orogenic activities in the Alpine--Carpathian realm. During Late Cretaceous times compression embraced the Inner Carpathians and several nappes with northward polarity developed. Subduction consumed the major part of the Pieniny Klippen Belt Ocean. Cherty limestones gave way to marls and flysch deposits. The Valais Ocean in the Alps finally closed (Froitzheim *et al*. 1996, Stampfli 1996, Golonka 2004). According to Froitzheim *et al*. (1996), the collision between the Austroalpine units and the Briançonnais terrane in the Alps started in the Early Paleocene.

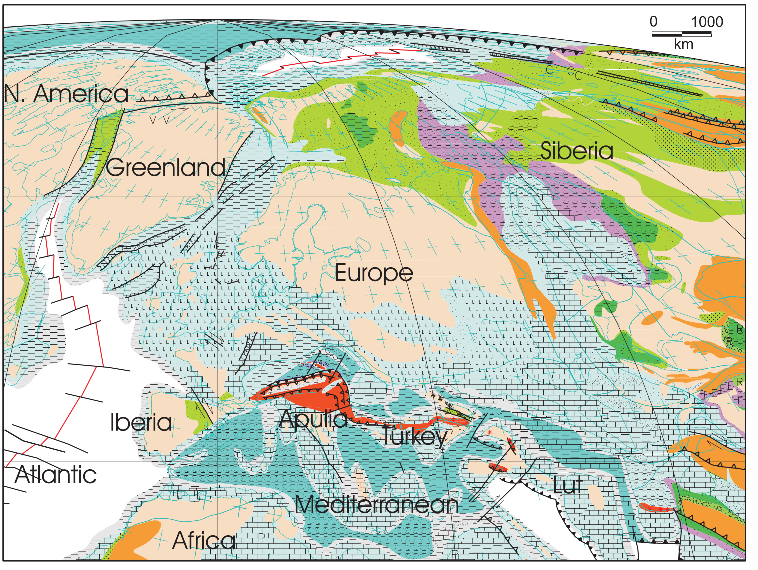

Fig. 4. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Late Cretaceous-Early Paleocene times

Fig. 4. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich w późnej kredzie-wczesnym paleocenie

As an effect of these movements the Inner Carpathians and Alps jointed with the Adria plate and the Alcapa terrane was created welding together of the Eastern Alps, Inner Carpathian, Tisa as well as smaller terranes, like the Bükk, Transdanubian or Getic (Golonka 2004). Several ridges have been uplifted as an effect of the orogenic process (Golonka *et al*. 2005a). The direction of these ridges was parallel to the uplifted zones in the European Platform (Fig. 4). From uplifted areas, situated within the Outer Carpathian realm as well as along its northern margin, enormous amount of clastic material was transported by various turbidity currents. This material filled the Outer Carpathian basins. Each basin had the specific type of clastic deposits, and sedimentation commenced in different time. Primary shortening events in the Balkans occurred in Bulgaria (Sinclair *et al*. 1997). The Vardar Ocean was closed during the Paleocene time (Şengör & Natalin 1996). Further eastwards the collision between Kirsehir, Sakariya and the Pontides (Yilmaz *et al*. 1997, Golonka 2004) closed the northern branch of Neotethys. The oceanic basins between Taurus and Kirsehir remained open. The northward movement of the Shatski terrane began closing of the proto-Black Sea (Kazmin 1997). The Lesser Caucasus approached the Transcaucasus and Talysh areas. At the end of the Cretaceous or in the Paleogene time, the Lesser Caucasus and perhaps Sanandaj-Sirjan and Makran plates were sutured to the Transcaucasus-Talysh-Southern Caspian-Lut system (Knipper & Sokolov 1974, Adamia 1991, Golonka 2004). According to Brunet *et al*. (2003), the subduction zone below the Lesser Caucasus was finished at the end of the Cretaceous, after the obduction of ophiolites in the Coniacian in the Sevan-Akera zone (Knipper *et al*. 2001).

In the northern Tethys area, the Maastrichtian was a period of widespread development of carbonate platform. Carbonate flysch was deposited in the basinal areas. At the end of the time slice, a drastic collapse of carbonate platforms occurred (Philip *et al*. 1996, Kiessling *et al*. 2003, Scheibner & Speijer 2008) due to increased tectonic activity, which resulted in regression or drowning. Compressional episodes occurred along the African-Arabian plate margin. These events included thrusting in the Moroccan High Atlas, folding of the Syrian arc, compression in the Sinai area (Moustafa & Khalil 1990, Guiraud & Bellion 1996, Frizon de Lamotte *et al*. 1998,) and inversion of the Central African Rift System (Golonka 2002).

The mantle plume attributed to the present-day Iceland hot spot was located between Baffin Island and Greenland (Lawver & Müller 1994, Golonka & Bocharova 2000, Ford & Golonka 2003). The mantle plume activity caused spreading of the Labrador Sea and rifting in the Baffin Bay (Gill *et al*. 1992, 1995, Holm *et al*. 1992, Larsen *et al*. 1992). Spreading in the Makarov Basin was perhaps also related to the opening of the Labrador Sea. The Eurekan orogeny, primarily a response to sea-floor spreading in the Labrador Sea and Baffin Bay, and rotation of Greenland affected the area between this plate and Canadian Arctic, mainly Ellesmere Island and the adjacent areas, developing a compressive foldbelt (Okulitch *et al*. 1998, Golonka 2000, von Gosen & Piepjohn 2003, Ford & Golonka 2003). A compressive foldbelt was developed in the westernmost Spitsbergen and North Greenland. Svalbard area and most of Barents Sea were uplifted (Fig. 4) (Rønnevik *et al*. 1982, Vorren *et al*. 1990, 1991, Skagen 1993, Bogatski *et al*. 1996, Ford & Golonka 2003, Golonka *et al*. 2003b, Cavanagh *et al*. 2006, Shipilov *et al*. 2006, Faleide *et al*. 2008). Compressional tectonics and volcanism is associated with the uplift. The Barents Sea was covered by continental clastics, including lacustrine fine-grained deposits. The fine-grained clastics prevailed on the margins of Labrador Sea, Makarov Basin and area of the future northern Atlantic. Rift developed between Greenland and northwestern Europe. Deep-water fine-grained clastics dominated sedimentation in the Vøring and Møre basins, offshore Norway (Doré 1991, Lundin & Doré 2002, Ford & Golonka 2003, Golonka *et al*. 2003b, Agterberg *et al*. 2007, Nøttvedt *et al*. 2008). Continental clastics, locally with coal were deposited in the West Siberian, Viluy and Chukotka basins, while marginal marine and shallow-marine, mainly fine-grained clastics existed in the Kara Sea area (Green *et al*. 1984, Bogdanov *et al*. 1998, Golonka *et al*. 2003b).

Dispersal of the continents and development of the passive margins and rift basins continued in the former Gondwana area (Lawver *et al*. 1985, Golonka 2000, 2002, Lawver & Gahagan 2003, Macdonald *et al*. 2003, Ghidella *et al*. 2007) (Figs 1, 5). Clastic sedimentation prevailed along partially uplifted margins of Africa, Antarctica and South America. Continental clastics filled basins of Africa (Bumby & Guiraud 2005). Development of the interior sag basins in South America, Africa, and Asia was associated with the renovation of ancestral failed rifts (Golonka 2000, 2002, Golonka *et al*. 2006). Large volcanic plateaus formed within the Indian Ocean due to mantle plumes activities (Coffin & Eldholm 1994, Charvis & Operto 1999, Dalziel *et al*. 2000, Golonka & Bocharova 2000, Coffin *et al*. 2002). Carbonate and mixed carbonate clastics were present in the area between Africa and Madagascar and on the Falkland (Malvinas) area (Ronov *et al*. 1989, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Wilson 2008, Scheibner & Speijer 2008).

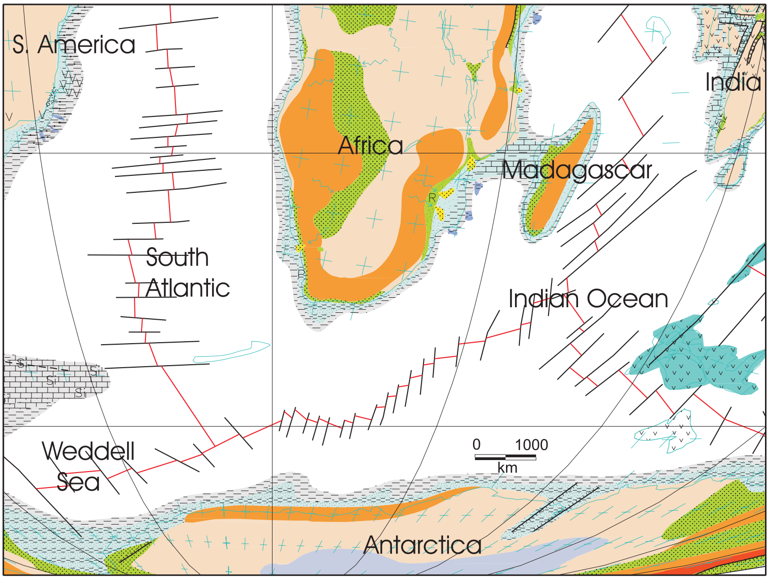

Fig. 5. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Late Cretaceous-Early Paleocene times

Fig. 5. Mapa tektoniki płyt, paleośrodowiska i litofacji południowego Atlantyku, południowo--zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich w późnej kredzie-wczesnym paleocenie

Late Paleocene-Middle Eocene

The Central, Equatorial and South Atlantic increased their widths, with westward movement of North and South America (Ladd 1977, Nürnberg & Müller 1991, Golonka 2000, 2002, Ford & Golonka 2003) (Figs 6, 7). Relatively narrow continental margins were established along Central Atlantic and part of the South Atlantic. Fine-grained clastics dominated sedimentation in the North American, Atlantic margin between Florida and Labrador Sea. The large Florida-Bahama carbonate platform, which was formed during Mesozoic times, continued throughout Paleogene (Ronov *et al*. 1989, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003). Carbonate sedimentation was extensive along low latitude, Western African margins and in some areas along the northeastern coast of South America. They were represented by transgressive carbonates consisting of coral reefs and related grainstones as well as by shelfal marls. The large Niger delta and fan was well developed during this time (Sahagian 1993, Ford & Golonka 2003, Bumby & Guiraud 2005, Deng *et al*. 2008, Olowokere 2008). The South Atlantic margins were typically narrow (except Argentina), and were dominated by fine-grained, shaly clastics with local development of deltas. Organic-rich sediments were deposited on the Trinidad, East Venezuela and Brazilian continental margins (Figueiredo & Milani 2000, Ford & Golonka 2003, Summa *et al*. 2003, Escalona & Mann 2006).

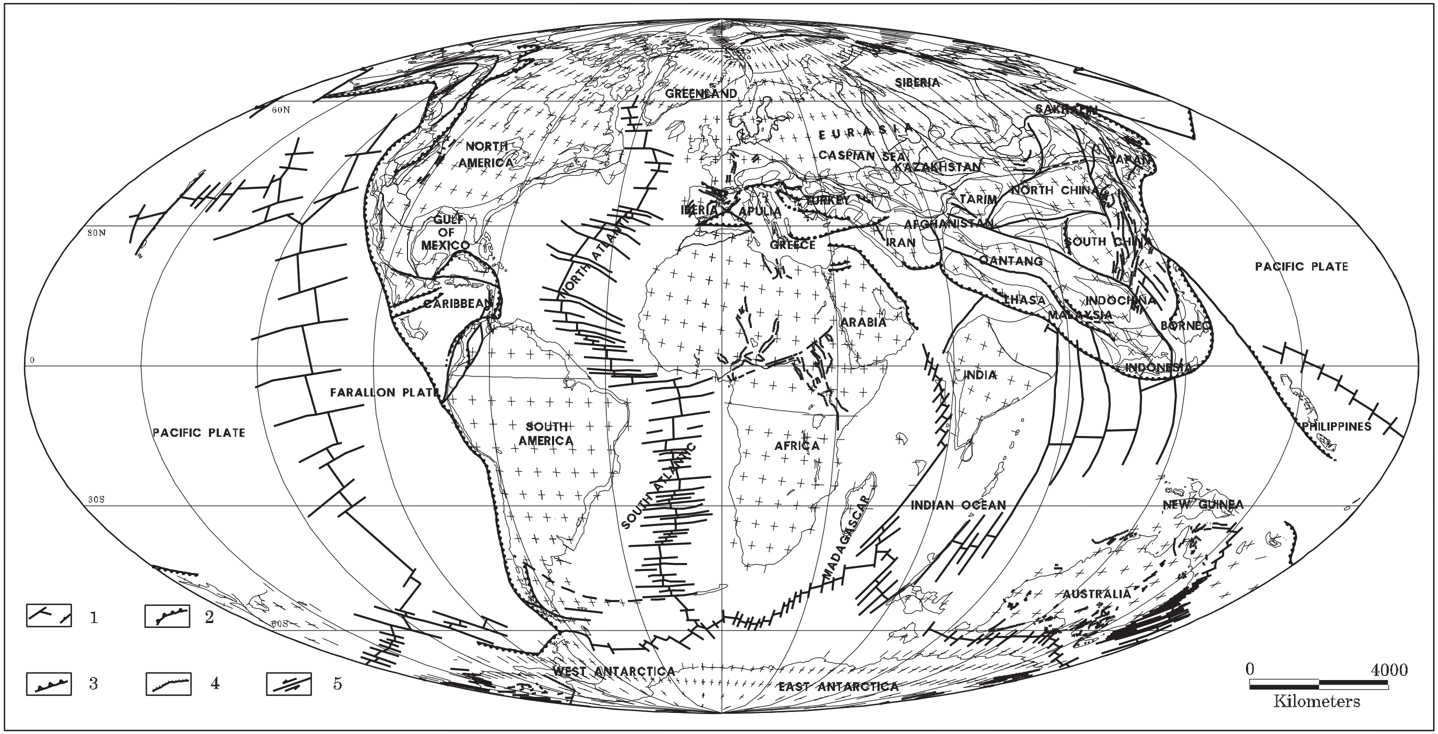

Fig. 6. Plate tectonic map of the Late Paleocene-Middle Eocene (plates position as of 53 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 6. Mapa tektoniki płyt wczesnego paleocenu-środkowego eocenu (pozycja płyt 53 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

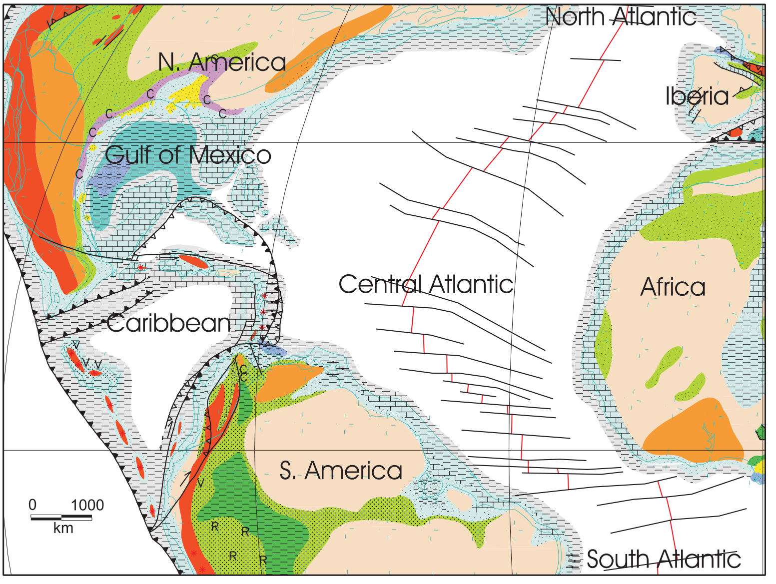

Fig. 7. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Late Paleocene-Middle Eocene times

Fig. 7. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich w późnym paleocenie-środkowym eocenie

South America moved westwards, encroaching on the Caribbean plate, creating volcanic arc and foredeep along the northwestern shelf (Pindell & Tabbutt 1995). Further development of the Panamanian arc took place along the western Caribbean plate margin (Burke 1988, Ross & Scotese 1988, Pindell & Barrett 1990, Golonka 2000, 2002, Golonka *et al*. 2006, Iturralde-Vinent 2006, Pindell *et al*. 2006, García-Gasco 2008). Further relative eastward movement of the Caribbean plate continued, with significant transpressive deformation along the northern and southern strike-slip boundaries (Guiraud & Bellion 1996, Acosta *et al*. 2007). Cuba had docked with North America. Yucatan was covered with shallow-marine carbonate platforms, while deep Gulf of Mexico was filled with deep-water clastics, fine-grained, locally with turbiditic fans (Ronov *et al*. 1989, Galloway & Williams 1991, Brewster-Wingard *et al*. 1997, Xue 1997, Watkins 1999, Galloway *et al*. 2000, Budd 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Rosenfeld 2005, Wilson 2008). Foreland basins along the Western Cordillera of North America (Sloss *et al*. 1960, Flynn 1986, Dickinson *et al*. 1988, DeCelles 2004, Carroll *et al*. 2006) were drained of marine seaways and become sites of fluvial/deltaic and lacustrine deposition (Fig. 7).

Central parts of the back-retro-arc basins in the Andes were inverted and uplifted. Clastic continental molasse-type of sedimentation with red beds, locally with volcanics and coals developed in the foredeep basins east of Andes (Pindell *et al*. 1991, Lamb *et al*. 1997, Olivero & Martinioni 2001, Golonka 2002, Ford & Golonka 2003, Hervouët *et al*. 2005, Duerto *et al*. 2006, Jordan *et al*. 2007). Continental clastics were also deposited within the Amazonian basins (Fig. 7).

The maps (Figs 6, 8) present the traditional (see Golonka *et al*. 1994, 2006, Golonka 2000, 2002, Kiessling *et al*. 2003) point of view about the onset of the collision of India with Asia occurred near the Paleocene-Eocene boundary (Gaetani & Garzanti 1991, Longley 1997). The Indian Ocean was formed, as a result of the northward movement of India (Royer & Sandwell 1989, Lawver *et al*. 1992, Golonka 2000, 2002). According to Searle (1996), this collision may have been diachronous, occurring earlier in northern Pakistan (60 Ma) then in Ladakh-southern Tibet. Recently, different points of view also exists and are presented for example by Ali and Aitchison (2008), who postulate collision time around 35 Ma. Northern part of India, including this part of Greater India now involved in the Himalayan structure, is covered by carbonates and mixed carbonates-clastics, while fine-grained clastic prevailed on the plate's southern passive margins with locally developed deltas (Fig. 8) (Kiessling *et al*. 2003, Acharyya 2007, Green *et al*. 2008). Carbonates, clastics, locally with volcanics occurred along the southern margin of Asia (Liu & Einsele 1994, Wan *et al*. 2002, Zhang *et al*. 2002, Golonka *et al*. 2006). Uplift condition continued within South China Indochina and Indonesia plates with continental deposits, red beds evaporites and volcanics (Golonka *et al*. 2006) (Fig. 8).

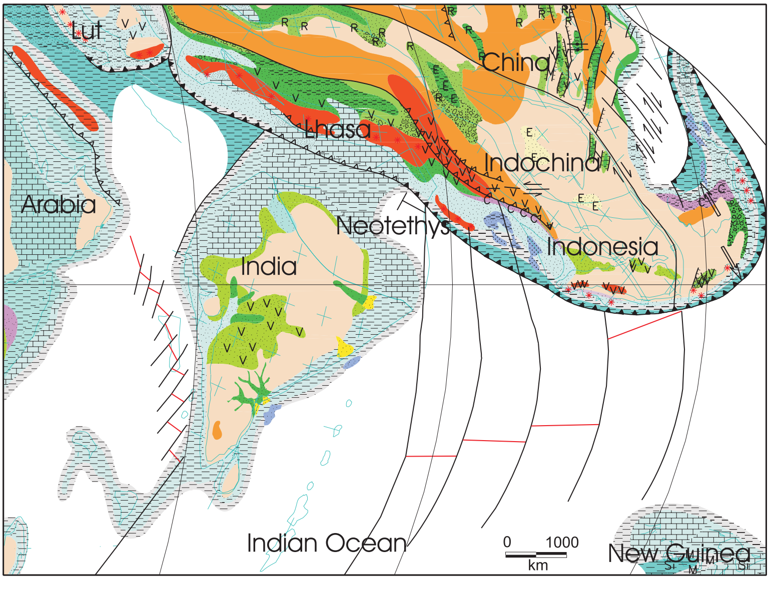

Fig. 8. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, eastern Neotethys and adjacent areas during Late Paleocene-Middle Eocene times

Fig. 8. Mapa tektoniki płyt, paleośrodowiska, i litofacji Azji południowo-wschodniej, wschodniej Tetydy oraz obszarów sąsiednich w późnym paleocenie-środkowym eocenie

Pull-apart basins and strike-slip faulting occurred in China (Golonka *et al*. 2006). Indochina perhaps initiated the movement southeastwards, with respect to South China along the left-lateral Red River Fault (Tapponnier *et al*. 1986, Lee & Lawver 1994, Golonka 2000, 2002, Golonka *et al*. 2006), the main stage of this movement occurred; however, at the later time the Red River Fault Zone in Yunnan, China and North Vietnam, up to 20 km wide, is one of the main strike-slip fault zones in SE Asia that separates the South China and Indochina blocks. In North Vietnam, the Red River Fault Zone is subdivided into three principal branches, up to 300 km long, orientated roughly NW-SE and named, from the NE to SW, the Lo River, Chay River, and Red River faults. These are mostly dextral and dextral-normal faults that show the southeastwards-increasing component of normal slip (Trinh 1995, Cuong & Zuchiewicz 2001). Between the Lo and Chay River faults recently growing anticlines, orientated WNW-ESE and W-E, have been found (Lacassin *et al*. 1993, 1994, Golonka *et al*. 2006).

Subduction of the Proto-South China Sea continued (Lee & Lawver 1994). Southward directed subduction resulted in crustal accretion along the northern margin of the Kalimantan and along the South Palawan Arc. A major strike-slip fault developed between the Indochina-Sumatra-plate and Borneo-Java (Longley 1997). The paleo-Ryukyu arc moved eastwards with the growth of the East China Sea Basin. The Philippine terranes (Seno & Maruyama 1984) appear for the first time on the Tejas I slice map. According to Longley (1997), during Eocene time, the motion of the Philippine plate changed, and a major transform fault developed into a subduction zone. According to Hall (1998, 2002), the Java--Sulavesi subduction system continued into the West Pacific beneath the eastern Philippines and Halmahera arc.

Back-arc extension in southeastern Asia led to rifting and seafloor spreading in the Celebes Sea Basin and the Makassar Basin, east of Borneo (Weissel 1980, Lee & Lawver 1994, Longley 1997). Rifting and stretching also occurred along the South Chinese platform. Back-arc extension and rapid subsidence continued in the East Chinese Sea Basin. Marine basins developed in the Philippines and in the proto-South China Sea (Seno & Maruyama 1984).

Carbonate platforms were present in the center of Arabia (Fig. 8), while fine-grained clastic prevailed in Arabian margin as well as in adjacent part of Somalia and on the Lut plate (Ronov *et al*. 1989, Dercourt *et al*. 1993, Guiraud & Bellion 1996, Kiessling *et al*. 2003, Golonka 2004, Guiraud *et al*. 2006, Scheibner & Speijer 2008, Wilson 2008). Northern African platforms exhibit rimmed margins that supply carbonate sediment to the adjacent basinal flysch wedges (Fig. 9).

The process of the closing of Neotethys by the Alpine orogeny continued. The Apulian plate was continuously moving northwards together with the Eastern Alpine (Austroalpine) and Inner Carpathian blocks. The Ligurian Ocean in the Alps was closed and subducted, and the closure of the Valais Ocean began (Froitzheim *et al*. 1996). Thrusting in the Eastern Alps was initiated in the Austroalpine upper plate, at about 55 Ma (Decker & Peresson, 1996). The closing of the Pieniny Klippen Belt basin in the Carpathians was also concluded (Golonka *et al*. 2000, Golonka & Picha 2006). Flysch, accumulated in the Alpine and Carpathian accretionary prisms and foreland basins in front of the moving northward plates. Carbonate platforms still existed on the Apulian (Adria) plate (Fig. 9) (Dercourt *et al*. 1993, Kiessling *et al*. 2003, Golonka 2004). The opening of new Paleogene basins proceeded at the expense of the Mesozoic oceanic basins. Some authors (e.g. Okay *et al*. 1996, Robinson *et al*. 1996) have suggested that the opening of the East Black Sea basin was connected with the simultaneous closure of the Great Caucasus basin.

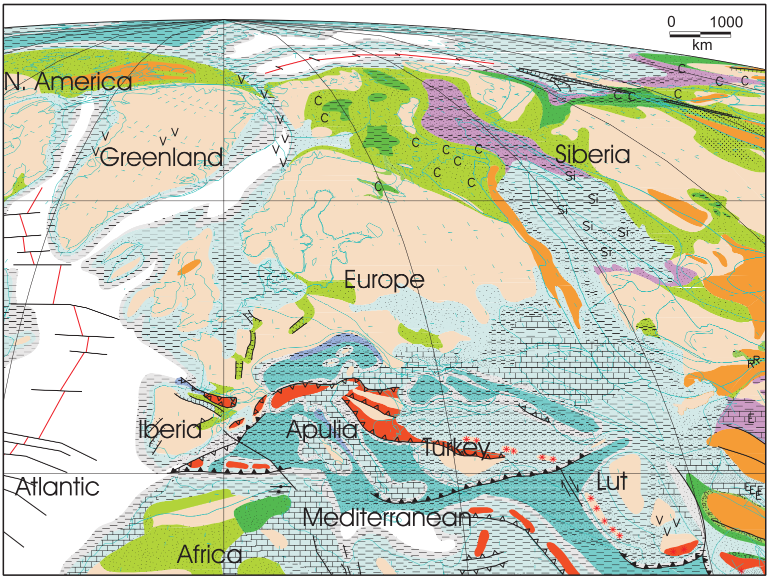

Fig. 9. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Late Paleocene-Middle Eocene time

Fig. 9. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich w późnym paleocenie-środkowym eocenie

The timing of the events is somewhat speculative (Golonka 2004). Perhaps the Jurassic--Cretaceous oceanic crust of the Greater Caucasus Basin was subducted under the overriding Scythian plate between the latest Cretaceous and the Late Eocene along with the opening of East Black Sea Basin (Fig. 9). The East Black Sea basin continued eastward to the Ajaro-Trialet and the Talysh basin (Adamia *et al*. 1974, Shcherba 1994, Banks *et al*. 1997, Kazmin 1997, Golonka 2004). In the Talysh basin in southern Azerbaijan, a thick volcanoclastic series with basalts, trachy-basalts, trachy-andesites and andesites followed by flysch with olistostromes. Altogether, Paleogene sediments reach a thickness 10 km (Gasanov 1992, 1996, Ali-Zade *et al*. 1996, Brunet *et al*. 2003). Vincent *et al*. (2005) connect Talysh extension with major extension and ocean spreading within the adjacent South Caspian Basin. Also Kazmin (1991, 1997) proposes the Eocene, while Berberian & Berberian (1981), Boulin (1991) and Abrams & Narimanov (1997), propose the Paleogene as the timing of the opening of the South Caspian Basin. Carbonates were deposited in Central Asia area (Ronov *et al*. 1989, Kiessling *et al*. 2003, Golonka 2004).

Most of the area of Western and Central Europe uplifted (Ziegler 1988, 1990, 1992, Baldschuhn *et al*. 1991, Mazur *et al*. 2005, Krzywiec 2006). North Sea and adjacent of continental Western Europe were covered by shallow sea with fine-grained sedimentation. (Fig. 9). Rhine Graben with continental clastics and volcanics originated at this time (Schäfer *et al*. 2005, van Balen *et al*. 2005, Schwarz & Henk 2005).

This onset of crustal separation in the initial phase of drifting and by the inception of seafloor spreading in the North Atlantic area was accompanied by extensive volcanism with seaward dipping basalts along the plate boundaries (Eldholm *et al*. 1990, Planke *et al*. 1991, Holm *et al*. 1992, Skogseid *et al*. 1992, Golonka 2000, 2002, Ford & Golonka 2003, Golonka *et al*. 2003b). A strong volcanic event resulted in nearly 1000 m of plateau basalts in the onshore East Greenland area. At the same time, dolerite sills and thin dikes were emplaced within the post-Devonian section throughout the region. Fine-grained clastics dominated sedimentation in the area between Europe and Greenland. At the same time, oceanic spreading was still active west of Greenland (Figs 6, 9). The opening of the Arctic Ocean (Eurasian Basin) was initiated, in the Late Paleocene (Kristoffersen 1990). It extended into the Laptev Sea are and onshore Siberia in the Moma rift area (Bogdanov *et al*. 1998, Franke *et al*. 2000). Svalbard and most of Barents Sea area were uplifted (Rønnevik *et al*. 1982, Eldholm *et al*. 1990, Vorren *et al*. 1990, 1991, Doré 1991, Johansen *et al*. 1993, Bogatski *et al*. 1996, Golonka *et al*. 2003b). The continental clastic sedimentation in the Barents and Timan-Pechora area included lacustrine deposits and coals (Fig. 9). Fine-grained clastic sedimentation dominated the West Siberian seaways (Vinogradov 1968, Ronov *et al*. 1989, Surkov *et al*. 1997, Golonka *et al*. 2003b, Akhmet'ev *et al*. 2004).

South Atlantic increased its widths, with westward movement of South America. Mixed clastics, sometimes with volcanics and organic-rich rocks were deposited on the African and South American margins (Fig. 10) (Ford & Golonka 2003).

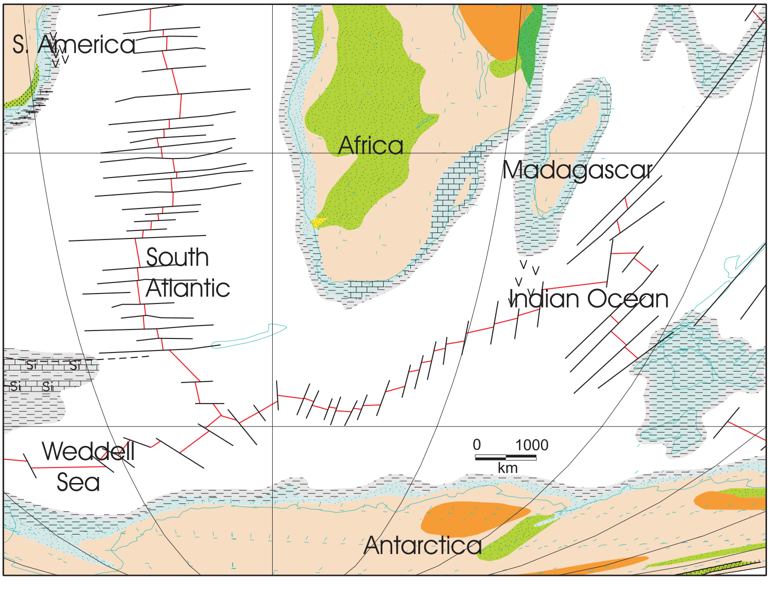

Fig. 10. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Late Paleocene-Middle Eocene times

Fig. 10. Mapa tektoniki płyt, paleośrodowiska i litofacji południowego Atlantyku, południowo-zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich w późnym paleocenie-środkowym eocenie

Sedimentation of continental clastics continued in the African basins (Bumby & Guiraud 2005). Fine-grained clastics and mixed carbonate clastics prevailed in the Falkland area (Ronov *et al*. 1989, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Scheibner & Speijer 2008, Wilson 2008). Seafloor spreading continued between Australia and Antarctica. The formation of the Indian Ocean was almost completed. Relatively deep-water clays and muds were deposited on the formerly placed seamounts and plateaus within the Indian Ocean around Madagascar, and slopes of Antarctica (Fig. 10) (Golonka 2000, 2002, Roberts *et al*. 2003, Das *et al*. 2007, Key *et al*. 2008).

Late Eocene

According to Pindell & Tabbutt (1995), the westward movement of South America across the mantle accelerated. This triggered the Incaic phase of the Andean tectonics, with a drastic uplifting of the mountain chain. Further, relative, eastward movement of the Caribbean plate continued, with significant transpressive deformation along the northern and southern strike-slip boundaries (Guiraud & Bellion 1996). Hispaniola moved along the strike-slip fault south of Cuba, eastwards, developing transpressive orogen (Fig. 10). East-west rifting began in the Cayman Trough. Fragments of Caribbeans were accreted to the northern part of South America with deformation and block rotation margin (Burke 1988, Ross & Scotes 1988, Pindell & Barrett 1990, Ave Lallemant 1997, Golonka 2000, 2002, Audemard & Audemard 2002, Golonka *et al*. 2006, Iturralde-Vinent 2006, Pindell *et al*. 2006, García-Gasco 2008). Sedimentation of continental clastics with red beds, continued in the large area east of Andes and within the Amazonian basins (Pindell *et al*. 1991, Lamb *et al*. 1997, Olivero & Martinioni 2001, Golonka 2002, Ford & Golonka 2003, Hervouët *et al*. 2005, Duerto *et al*. 2006, Jordan *et al*. 2007). Marine clastics were present in Venezuela (Ford & Golonka 2003, Summa *et al*. 2003). Large carbonate platforms continued on Yucatan and on Bahama platform, while deep Gulf of Mexico was filled with deep-water clastics rimmed on the northern margin with marginal marine strata with coals (Ronov *et al*. 1989, Galloway & Williams 1991, Brewster-Wingard *et al*. 1997, Xue 1997, Watkins 1999, Galloway *et al*. 2000, Budd 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Rosenfeld 2005, Wilson 2008).

Inversion of the High Atlas Mountains in Morocco marked a new, convergent boundary between stable Africa and the Moroccan and Oran mesetas (Ricou 1996). A major compressive event took place at the Middle-Late Eocene transition (Guiraud & Bellion 1996).

The northern African platforms exhibit rimmed margins that supply carbonate sediment to adjacent basinal flysch wedges. Interiors of Northern Africa were filled with continental clastics (Fig. 12) (Bumby & Guiraud 2005).

The collision of India and Eurasia continued (Figs 11, 13). Metamorphism and crustal thickening reached a peak about 40 Ma in northern Pakistan, propagating later southward (Searle 1996). Oceanic subduction ceased beneath the Indian-Eurasian collision zone (Longley 1997). India's interior was uplifted, locally with continental clastics and volcanics. Numerous delta and sea-fans developed along the plate margins.

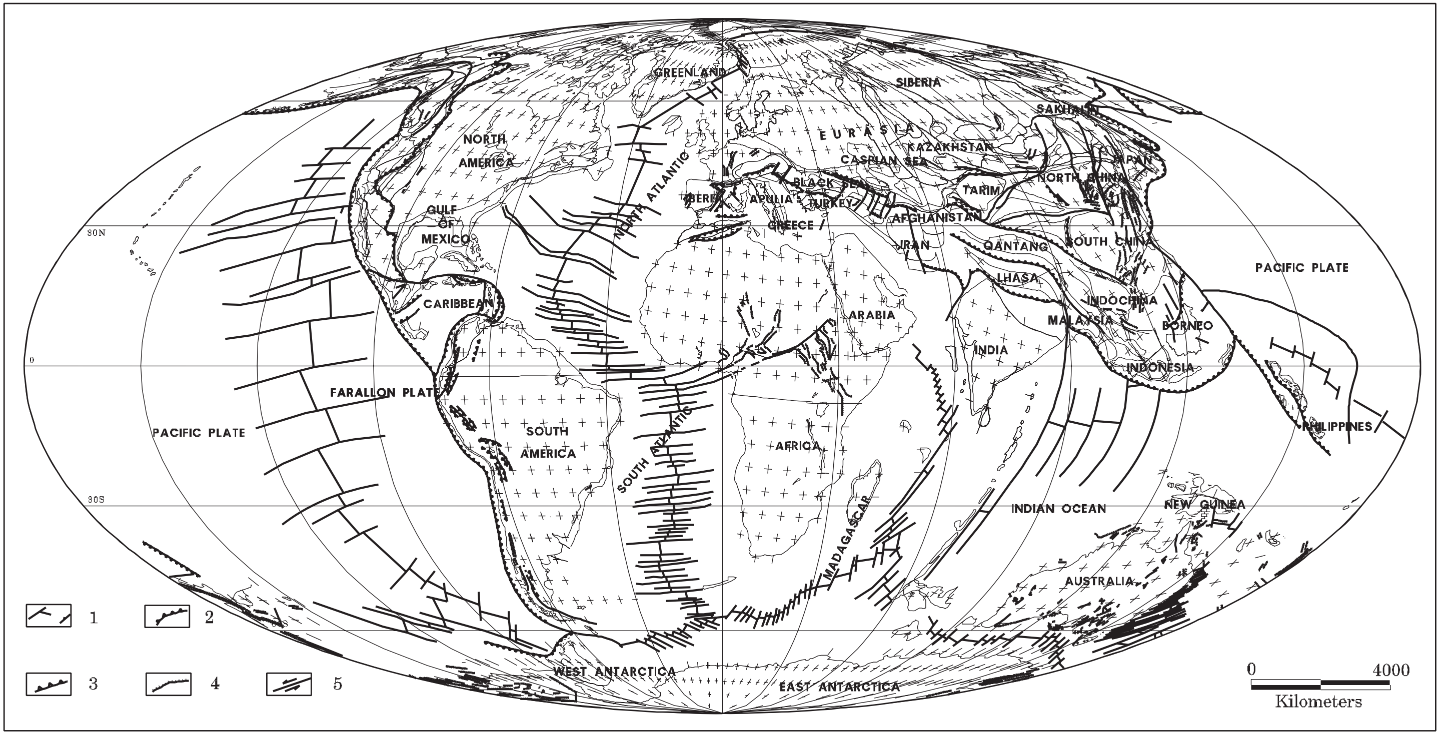

Fig. 11. Plate tectonic map of the Late Eocene (plates position as of 45 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 11. Mapa tektoniki płyt późnego eocenu (pozycja płyt 45 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

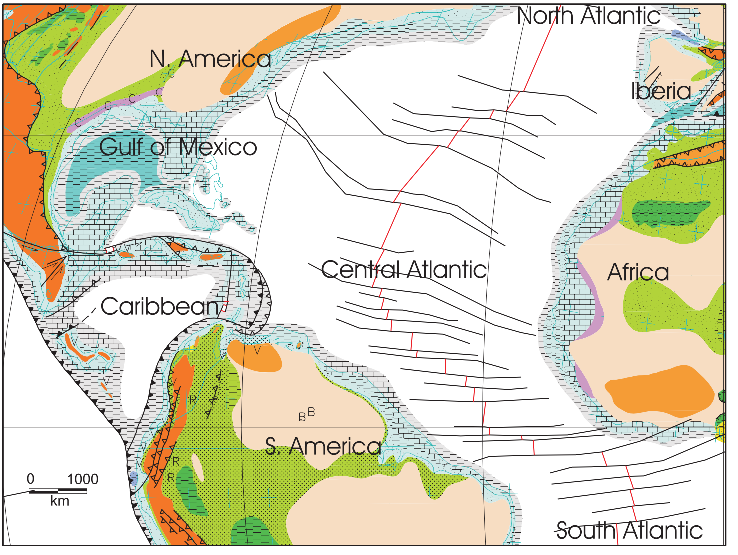

Fig. 12. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Late Eocene times

Fig. 12. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich w późnym eocenie

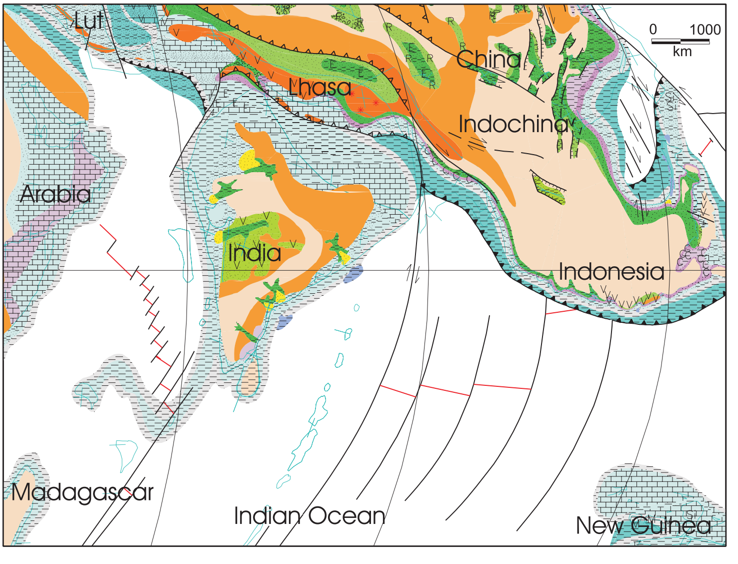

Fig. 13. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, Indian Ocean and adjacent areas during Late Eocene times

Fig. 13. Mapa tektoniki płyt, paleośrodowiska i litofacji Azji południowo-wschodniej, Oceanu Indyjskiego oraz obszarów sąsiednich w późnym eocenie

The India plate acted as an inventor on Asia causing southeastward and southward extrusion of Indochina and South China plates (Golonka *et al*. 2006). The movements of extruded plates initiated major strike slip faults in Southeast Asia. The Red River Fault Zone activity was approaching its main phase. According to Jolivet *et al*. (2001), the Red River Fault Zone is rooted in an horizontal shear zone at the brittle/ductile transition separating the upper and middle crust from the lower crust, and the sinistral strike-slip motion was first transpressional (?40–25 Ma), and then transtensional, leading to fast exhumation from 24 to 17 Ma. According to Wang *et al*. (2000), transpressional tectonics in the Red River Fault Zone may have started as early as ~ 42 Ma. Their geochronological dates reveal two distinctive magmatic episodes in eastern Tibet and Indochina: one between 42 and 24 Ma and another since 16 Ma. The older volcanic rocks and minor intrusions are distributed along the entire length of the Red River Fault Zone and its northern extension. According to Lan *et al*. (2000) around 40 Ma the extension resulting from the India-Asia collision caused a mantle input to mix with recycled sediments generating granitic rocks. According to Wang *et al*. (2000), the rocks of earlier igneous activity (42–24 Ma) include syenite, trachite, shoshonitic lamprophyre and basaltic trachyandesite. The strike-slip movement is related to the origin of pull-apart basins with continental clastic deposition (Fig. 13).

The old ridge began to be subducted beneath the Sunda Arc (Indonesia). Rifting in the Celebes Sea continued (Lee & Lawver 1994). East China Sea reached its maximum size (Golonka *et al*. 2006). Subduction of the proto-South China Sea continued (Lee & Lawver 1994). According to Hall (1998, 2002), the opening of the West Philippine-Celebes Sea Basin required the subduction of the proto-South China Sea beneath Luzon and Sulu arc. The Australasian margin collided with New Guinea arc resulting in emplacement of New Caledonia ophiolite and beginning of subduction beneath Papua-New Guinea.

Shallow-water carbonates developed on the Arabian plate (Fig. 13), providing good reservoirs for hydrocarbon exploration (Whittle & Alsharhan 1994, Whittle *et al*. 1996, Alsharhan & Nairn 1997, Dull 2005, El-Saiy & Jordan 2007). Widespread distribution of carbonates occurred in the central and eastern parts of North Africa (Ziegler 1988, Dercourt *et al*. 1993, Philip *et al*. 1996, Macgregor & Moody 1998, Bolle *et al*. 1999, Bolle 2000, Bensalen 2002, Kiessling *et al*. 2003, Guiraud *et al*. 2005, Shiref & Salaj 2007, Scheibner & Speijer 2008). Carbonate platforms continued to exist on the Apulian/Adrian plate, but, in the remaining part of the northern Tethys were replaced by a clastic sedimentation (Fig. 14).

During the Eocene time, Lesser Caucasus, Sanandaj-Sirjan and Makran plates were sutured to the Transcaucasus-Talesh-Southern Caspian-Lut system (Adamia 1991, Golonka 2004, 2007a). The subduction zone was locked and jumped to the Scythian-Turan margin. The western segment of this subduction was located along the northern margin of the Eastern Black Sea, on the Greater Caucasus area and south off the Apsheron Peninsula and ridge. The major transform fault system in the Western Turkmenistan basin area separated the eastern and western segments of subduction. This fault system is buried deeply below West Turkmen Basin Neogene sediments. The eastern segment was located along the South Kopet Dagh margin approximately 200–300 km south of the Apsheron ridge.

The subduction jump produced the trench pulling force, which influenced all plates between Black Sea and Sistan Ocean in Afghanistan. The timing of movement initiation and movement velocity is different for different plates. This difference caused the origin of several major strike slip faults of SW-NE orientation, which cut both continental crust and Jurassic-Cretaceous oceanic crust (Jackson 1992, Kopp 1997, Golonka 2007a). The most important are Araks fault, which separates the Lesser Caucasus block and Transcaucasus block from the Talesh plate, and the Lahijan fault within Alborz Mountains. The extension of the Lahijan fault, which separated the South Caspian Microcontinent (SCM) from South-West Caspian Basin, is buried deeply below the South Caspian Neogene sediments. The northward movement of the South Caspian Microcontinent resulted in rifting between the SCM and Alborz plates. The Jurassic-Cretaceous oceanic crust of the Eastern Black Sea – Greater Caucasus Basin was subducted under the overriding Scythian plate. The northward movement of the Shatski Rise block caused opening of the Eastern Black Sea (Robinson *et al*. 1996, Banks *et al*. 1997). The northward movement of the Transcaucasus block caused the collision and formation of the Greater Caucasus orogenic belt (Adamia 1991). A marine environment spread throughout the Scythian-Turan platform and adjacent areas during-Eocene time, which was characterized by mixed deposition of clastic and carbonate rocks (Fig. 14).

The Paleocene inversion in the Carpathians was followed by a new episode of subsidence, which accelerated during the Lutetian and Priabonian (Oszczypko *et al*. 2003, Golonka & Picha 2006). The accretionary prism was building gradually, causing the northward migration of depocenters. Thin-bedded flysch deposits passed into a thick complex of turbidites and fluxoturbidites. Foreland basin development proceeded in southern Europe, coinciding with a general uplift of the European continent. The closure of the Pindos Ocean began (Robertson *et al*. 1991). Compression continued in the Balkan area in Bulgaria (Tari *et al*. 1997).

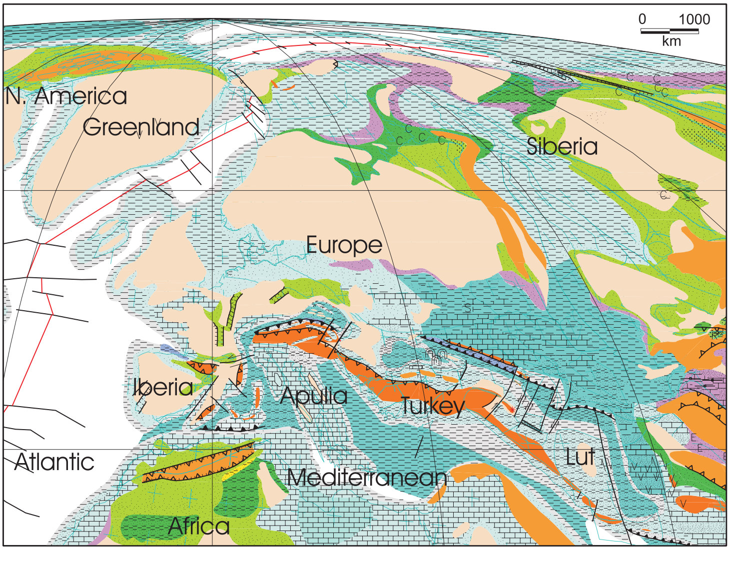

Fig. 14. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Late Eocene times

Fig. 14. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich w późnym eocenie

Seafloor spreading finally shifted from western to eastern Greenland and the North Atlantic (Figs 11, 14). The opening of the North Atlantic was linked to mantle plume, associated with the Iceland hot spot as postulated by White & McKenzie (1989). According to White (1992), shortly after the Iceland plume was reactivated, the extension between Greenland and the north-western European margin continued until its development into a full oceanic spreading centre. The voluminous volcanic complexes, containing wedges of seaward-dipping reflectors, were deposited in the vicinity of the continental-oceanic transition (Planke *et al*. 1991, Skogseid *et al*. 1992, Coffin & Eldholm 1994). Clastics, mainly fine-grained, locally with volcanics were deposited along North Atlantic and (Ziegler 1988, 1990, Joy 1992, Bull & Masson 1996, Ford & Golonka 2003, Golonka *et al*. 2003b, Laberg *et al*. 2005, Agterberg *et al*. 2007, Rasmussen *et al*. 2008). Svalbard was uplifted, but shallow water environment prevailed on large part of Barents Sea (Rønnevik *et al*. 1982, Vorren *et al*. 1990, 1991, Johansen *et al*. 1993, Skagen 1993, Bogatski *et al*. 1996, Musatov & Pogrebitskij 2000, Golonka *et al*. 2003b, Shipilov *et al*. 2006, Rasmussen *et al*. 2008). The continental clastic sedimentation prevailed Timan-Pechora area and adjacent part of South Barents Sea included lacustrine deposits and coals (Fig. 14). Fine-grained clastic sedimentation dominated the West Siberian seaways (Vinogradov 1968, Ronov *et al*. 1989, Volkova & Kul'kova 1996, Kulkova & Volkova 1997, Surkov *et al*. 1997, Golonka *et al*. 2003b, Akhmet'ev *et al*. 2004). Continental clastic, fluvial and lacustrine; with coals, were deposited it the other, smaller Siberian basins Vinogradov 1968, Ronov *et al*. 1989, Zonenshain *et al*. 1990, Parfenov 1991, Martinson 1998, Paech *et al*. 2000, Parfenov *et al*. 2001). Lignite was especially present in the Moma basin.

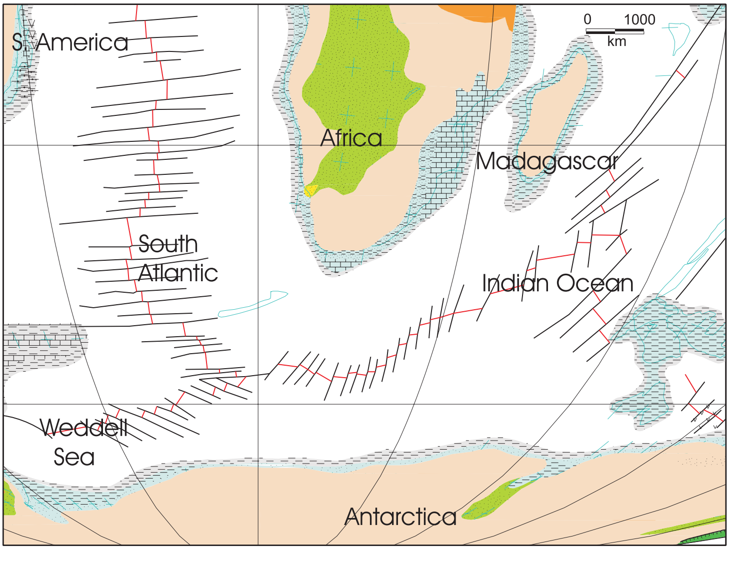

Fig. 15. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Late Eocene times

Fig. 15. Mapa tektoniki płyt, paleośrodowiska i litofacji południowego Atlantyku, południowo--zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich w późnym eocenie

The South Atlantic increased its width during Eocene times (Figs 11, 15). Mixed clastics, sometimes with volcanics and organic-rich rocks were deposited on the Atlantic margins in Africa and South America (Fig. 15) (Ford & Golonka 2003). Delta of the Orange River supplied coarse clastics (Bluck *et al*. 2007). Sedimentation of continental clastics continued in the African basins (Bumby & Guiraud 2005). Fine-grained clastics and mixed carbonate clastics prevailed in the Falkland area (Ronov *et al*. 1989, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Scheibner & Speijer 2008, Wilson 2008). Seafloor spreading continued between Australia and Antarctica. Relatively deep-water clays and muds were deposited on the formerly placed seamount and plateaus within the Indian Ocean around Madagascar, and slopes of Antarctica (Fig. 15) (Golonka 2000, 2002, Roberts *et al*. 2003, Das *et al*. 2007, Key *et al*. 2008). Shallow-marine and carbonate clastic sedimentation occurred along the Mozambique Channel Indian Ocean margins and offshore South Africa (Förster 1975, Salman & Abdula 1995, Ford & Golonka 2003, Kiessling *et al*. 2004, Key *et al*. 2008).

Oligocene

Atlantic Ocean continued its spreading during Oligocene times (Fig. 16). The development of the eastern Caribbean island arc occurred (Fig. 17), while the Panamanian arc nearly collided with South America (Burke 1988, Ross & Scotese 1988, Pindell & Barrett 1990, Ave Lallemant 1997, Golonka 2000, 2002, Audemard & Audemard 2002, Golonka *et al*. 2006, Iturralde-Vinent 2006, Pindell *et al*. 2006, García-Gasco 2008). According to Coates *et al*. (2004), precollisional arc-related rocks of Panama consisted of 4000 m of pillow basalts and volcanoclastics, and biogenic calcareous and siliceous deepwater sediments. Left-lateral strike-slip motion was active between the North America and Caribbean plates and deformation took place under a transpressive tectonic regime. A collisional unit developed on Hispaniola, which was located south of Cuba (Pérez-Estaún *et al*. 2007). Andean folding and thrusting continued (Lamb *et al*. 1997).

Continental clastics, locally with coals and organic-rich source rocks, were deposited over a large area east of Andes and within the Amazonian basins (Pindell *et al*. 1991, Lamb *et al*. 1997, Olivero & Martinioni 2001, Golonka 2002, Ford & Golonka 2003, Hervouët *et al*. 2005, Duerto *et al*. 2006, Jordan *et al*. 2007). Marine clastics with source rocks were deposited in Venezuela (Ford & Golonka 2003, Summa *et al*. 2003). Carbonate platforms, sometimes with reefs partially covered Greater Antilles Islands (Ford & Golonka 2003, Kiessling *et al*. 2003, Baron-Szabo 2005, Johnson & Pérez 2006). According to Mutti *et al*. (2005), carbonate “megabank” extended from the Honduras/Nicaraguan mainland to the modern island of Jamaica. Large carbonate platforms continued on Yucatan and on Bahama platform, as in previous time slice while deep Gulf of Mexico was still filled with deep-water clastics rimmed on the northern margin with marginal marine strata with coals (Ronov *et al*. 1989, Galloway & Williams 1991, Brewster-Wingard *et al*. 1997, Xue 1997, Watkins 1999, Galloway *et al*. 2000, Budd 2002, Ford & Golonka 2003, Kiessling *et al*. 2004, Rosenfeld 2005, Wilson 2008).

Fig. 16. Plate tectonic map of the Oligocene (plates position as of 36 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 16. Mapa tektoniki płyt oligocenu (pozycja płyt 36 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

According to Frizon de Lamotte *et al*. (2009 in press), the general inversion and orogenesis of the Atlas System occurred during two distinct episodes, Middle-Late Eocene-Oligocene and Late Miocene-Pliocene, respectively, whereas during the intervening period, the Africa-Europe convergence was mainly accommodated in the Rif-Tell system. According to Summerhayes *et al*. (1981), the Oligocene earth movements gave rise more or less to the present shelf-slope configuration. Coarse clastics, fine-grained clastics and locally carbonates were deposited on the Northern African margins, while interiors of Northern Africa were filled with continental clastics, mainly sands and gravel mounds (Fig. 17) (Said 1983, Coward & Ries 2003, Ford & Golonka 2003, Bumby & Guiraud 2005, Swezey 2009).

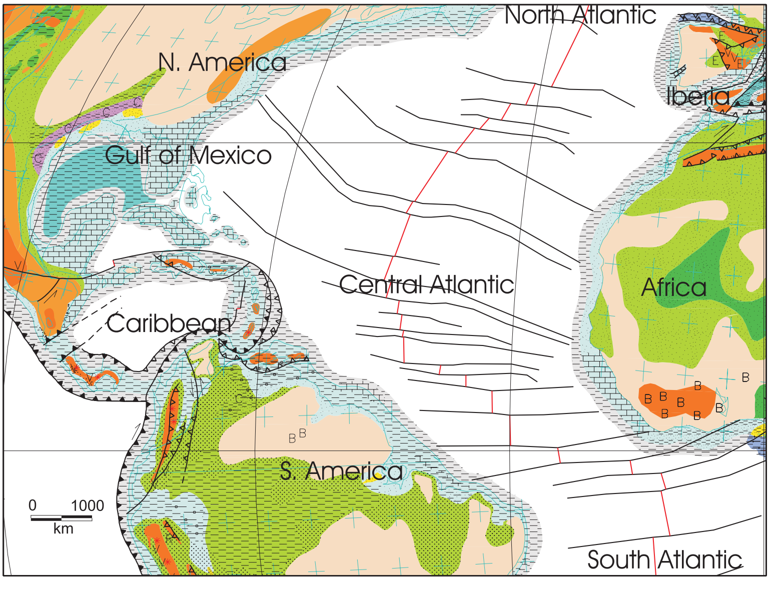

Fig. 17. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Oligocene times

Fig. 17. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich w oligocenie

The collision of India and Eurasia continued (Figs 16, 18). Metamorphism and crustal thickening reached their peak in the Zanskar area (Searle 1996, Golonka *et al*. 2006). The development of the molasse basins continued in the Himalayan belt foreland (Burbank *et al*. 1996). Marine clastics, locally carbonates as well as marginal and terrestrial clastics sometimes with coals were deposited there. Like during the previous time slice India's interior was uplifted, locally with continental clastics and volcanics. Numerous delta and sea-fans developed along the plate margins. Collision and suturing of India to Asia caused extensive strike-slip faulting in Asia (Kopp 1997). According to Hall (1998, 2002), the Indian Ocean subduction continued at the Sunda-Java trenches, and also beneath the arc extending from Sulawesi though the east Philippine to Halmaheru. Active volcanism and subduction took place in the Philippine island arc. The closing of the proto-South China Sea continued

(Lee & Lawver 1994). Inversion and folding in East China and the first phase of formation of the Taiwan-Sinzi Folded zone was caused by a collision between the paleo-Ryukyu Arc and the northern part of the Philippine Sea plate (Kong 1998, Kong *et al*. 2000). The Luconian terrane collided with NW Borneo (Longley 1997).

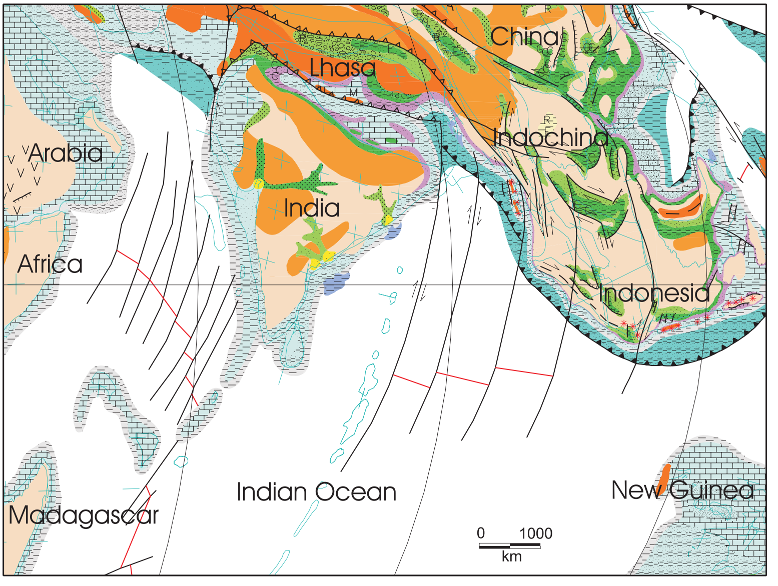

Fig. 18. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, Indian Ocean and adjacent areas during Oligocene times

Fig. 18. Mapa tektoniki płyt, paleośrodowiska i litofacji Azji południowo-wschodniej, Oceanu Indyjskiego oraz obszarów sąsiednich w oligocenie

Sedimentation of shallow-water carbonates continued on the Arabian plate (Figs 18, 19) (Whittle & Alsharhan 1994, Whittle *et al*. 1996, Alsharhan & Nairn 1997, Dull 2005, El-Saiy & Jordan 2007). The Sanandaj-Sirjan plate began to thrust over the Arabian Platform forming the Zagros Mountains (Berberian & Berberian 1981, Dercourt *et al*. 1993, Golonka 2004, Fakhari *et al*. 2008, Heydari 2008). The main cause of thrusting in the Zagros Mountains (Figs 18, 19), according to Şengör & Natalin (1996), was the counterclockwise rotation of the Arabian plate. Collision of the Lut block with the Turan platform in Central Asia caused the Kopet Dagh foldbelt to form (Kopp 1997).

The subduction zone beneath the Scythian-Turan margin of Eurasia (Sobornov 1994) produced a trench-pull force which caused northward movement of the plates between the Black Sea and Afghanistan (Fig. 19), closure of the Greater Caucasus, Sebzevar and Sistan Oceans and reorganization of the South Caspian Sea (Golonka 2004, 2007a). The Sistan Ocean was closed in eastern Iran, between Helmand and Lut plates (Sengör & Natalin 1996). Collisions continued in the area between Africa and Eurasia. The conclusion of the compression of the Balkanides, in Bulgaria, occurred during the Oligocene time (Sinclair *et al*. 1997). The Pindos Ocean was finally closed (Robertson *et al*. 1991). The collision of

Apulia as well as the Alpine-Carpathian terranes with the European plate continued (Decker & Peresson 1996, Frasheri *et al*. 1996). The metamorphism of the undercrusted Penninic nappes in the Alps reached peak thermal conditions at about 30 Ma (Kurz *et al*. 1996). The Calabrian terranes in the Western Mediterranean began to progress eastward (van Dijk & Okkes 1991).

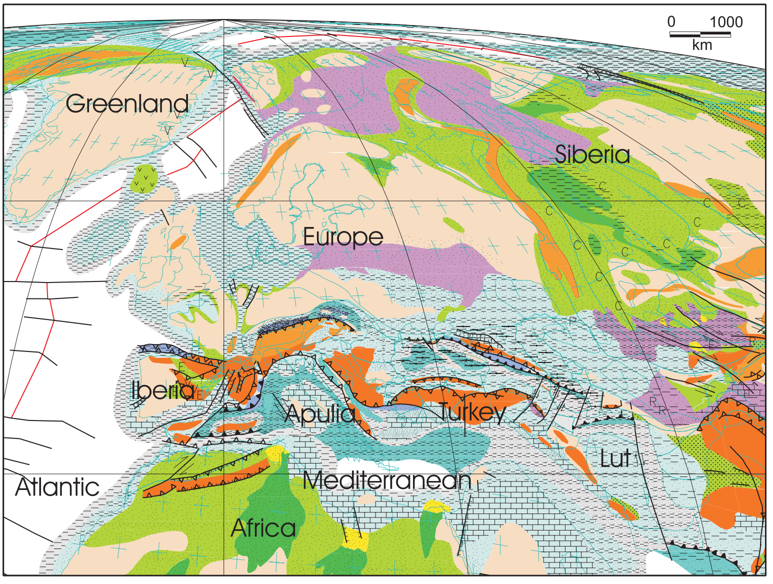

Fig. 19. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Oligocene times

Fig. 19. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich w oligocenie

During the Eocene-Oligocene times, the Paratethys Sea developed in Europe and central Asia, ahead of the progressing northwards orogenic belts (Dercourt *et al*. 1993, Popov *et al*. 1993). The main part of Neotethys was closed. The Neotethys remnants, foreland basin of the Alpine orogens, and reorganized Greater Caucasus-Caspian basin with the adjacent parts of the Scythian-Turan platform formed the Paratethys Sea. Large area was covered by shallow sea with prevailing mixed coarse and fine clastic deposition (Fig. 19). The Paratethys was isolated from the world ocean. This isolation and persistent low pressure system during Oligocene-Early Miocene time (Golonka 2003, 2007a) generated favorable condition for deposition and preservation of the organic-rich shales. The Maykop formation containing several layers of organic-rich shales was deposited in north of Greater Caucasus, in the Terek-Caspian Basin, Kura Basin, and perhaps in the parts of South Caspian Basin (Popov *et al*. 1993, Abrams & Narimanov 1997, Inan *et al*. 1997, Devlin *et al*. 1999, Devlin 2000, Brunet *et al*. 2003, Golonka 2007a, Stolyarov & Ivleva 2007, Hudson *et al*. 2008). The South Caspian Microcontinent was probably emerged during the Oligocene--Early Miocene and received a minimum amount of sediments, so we can expect the absence of the Maykop formation in this area.

Continental clastics, locally with coal, were deposited in the West Siberian, Viluy and Chukotka basins, while marginal marine environment existed in the Kara Sea area (Fig. 19) (Green *et al*. 1984, Ronov *et al*. 1989, Volkova & Kul'kova 1996, Kulkova & Volkova, 1997, Bogdanov *et al*. 1998, Akhmet'ev *et al*. 2001, Golonka *et al*. 2003b, Kuz'mina & Volkova 2008).

The geodynamic evolution of the basins in the Alpine-Carpathian belt led to a transition from flysch to molasse type of sedimentation (Fig. 19). Rifting events were initialized during the Oligocene time in the several countries in Europe between France and Ukraine (Rutkowski 1986, Ziegler 1988, 1990, 1992, Żytko *et al*. 1989, Wilson & Downes 1991, Bois 1993, Wilson 1994) and were associated with the alkaline volcanism. According to Bois (1993), extension occurred in part of the European plate, with the rifting of the Rhine, Limagne and Bresse Trough, contemporaneous of the climax of Alpine compression. Part of this rift system included the Gulf of Lions, associated with the mantle plume, expressed by volcanics in the Massif Central and Provence and on Corsica and Sardinia (Wilson & Downes 1991). Rifting in this area was followed by oceanic seafloor spreading and drifting of the Corsican and Sardinian plates (Bois 1993, Ricou 1996). The North Sea subsidence, renewed during the Tertiary (Joy 1992), could have been related to Central European rifting. This area was covered with fine-grained clastics (Berstad & Dypvik 1982, Ziegler 1988, 1990, 1992, Jordt *et al*. 1995, 2000, Eidvin & Rundberg 2007, Japsen *et al*. 2008, Rasmussen *et al*. 2008, Marcussen *et al*. 2009).

Seafloor spreading continued in the North Atlantic, with further opening of the Eurasian Basin in the Arctic (Kristoffersen 1990, Zonenshain *et al*. 1990). This basin was separated from the Canadian basin by the Lomonosov Ridge.

Svalbard moved away from Greenland (Livsic 1992, Manby & Lyberis 1996) along right-lateral strike slip transforming the previously sheared margin into a passive margin. The West Svalbard orogenic belts are considered by Lyberis & Manby (1999a) as strike--slip or transpressive orogen resulting from the continental collision and lateral escape. En-echelon folds, flower structures, and local extensional features were interpreted as evidence for transpressive deformation. The compressional aspects of the orogenic belt have also been emphasized (Golonka 2000, 2002). The Central Spitsbergen Basin area developed as a foreland at this time.

Mature seafloor spreading in the Southern Hemisphere led to northward movement of the continents. Rapid spreading continued between Australia and Antarctica (Lawver & Gahagan 1993). Formation of the Scotia Sea plate reached the Drake Passage, which allowed the circum-Antarctic seaway to develop (Lawver *et al*. 1992, Macdonald *et al*. 2003). Mixed clastics were deposited on the South American and Antarctic Margins with addition of carbonates in the Falkland area (Fig. 20) (Ronov *et al*. 1989, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Scheibner & Speijer 2008, Wilson 2008). Fine-grained and mixed clastic dominated along African Margins, while sedimentation of continental clastics continued within the African basins (Bumby & Guiraud 2005). Carbonate occurrences took place on Madagascar (Golonka 2000, 2002). Seafloor spreading continued between Australia and Antarctica.

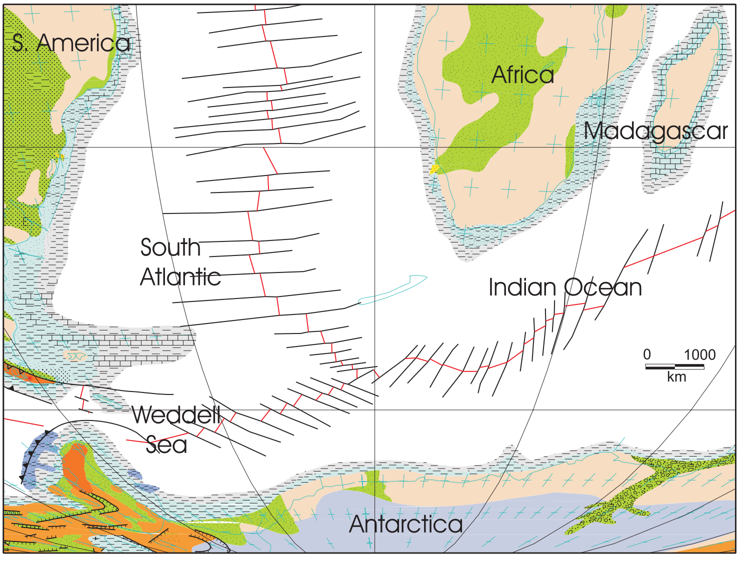

Fig. 20. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Oligocene times

Fig. 20. Mapa tektoniki płyt, paleośrodowiska i litofacji południowego Atlantyku, południowo--zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich w oligocenie

A major cooling trend took place, with highly variable continental climates, associated with latitudinal and orographic effects. An ice sheet formed on the southern hemisphere (Fig. 20) (Golonka 2000, 2002, Pfuhl & McCave 2005, Golonka *et al*. 2006, Ivany *et al*. 2006, Pekar *et al.* 2006, Barker *et al*. 2007, DeConto *et al*. 2008, Francis *et al*. 2008, Siegert *et al*. 2008). According to Abreu & Baum (1997), the isotope curve indicates that the ice sheet on Antarctica experienced phases of growth during the Late Eocene to Early Oligocene, followed by a decrease in volume in the Late Oligocene. The onset of Antarctica glaciation was related to the opening of seaway around Antarctica (Lawver *et al*. 1992, Golonka 2000, 2002, Golonka *et al*. 2006).

Early Miocene

This was a time of mature seafloor spreading globally, with local rifting events as well as of continental collisions in the Alpine-Himalayan belt (Fig. 21). Central Atlantic Ocean continued its spreading. Subduction and orogenesis continued along the entire Cordillera of North and South America (Figs 21, 22). The rate of motion of South America, relative to the mantle, slowed during the Early Miocene times (Pindell & Tabbutt 1995). The Antilles arc continued its eastward movement. The Panamanian Isthmus was established with still open seaways. Strike-slip fault was active between Cuba and Hispaniola plates causing transpressive deformations (Burke 1988, Ross & Scotese 1988, Pindell & Barrett 1990, Ave Lallemant 1997, Golonka 2000, 2002, Audemard & Audemard 2002, Golonka *et al*. 2006, Iturralde-Vinent 2006, Pindell *et al*. 2006, García-Gasco 2008). Another strike-slip fault was located between Caribbean and South American plates.

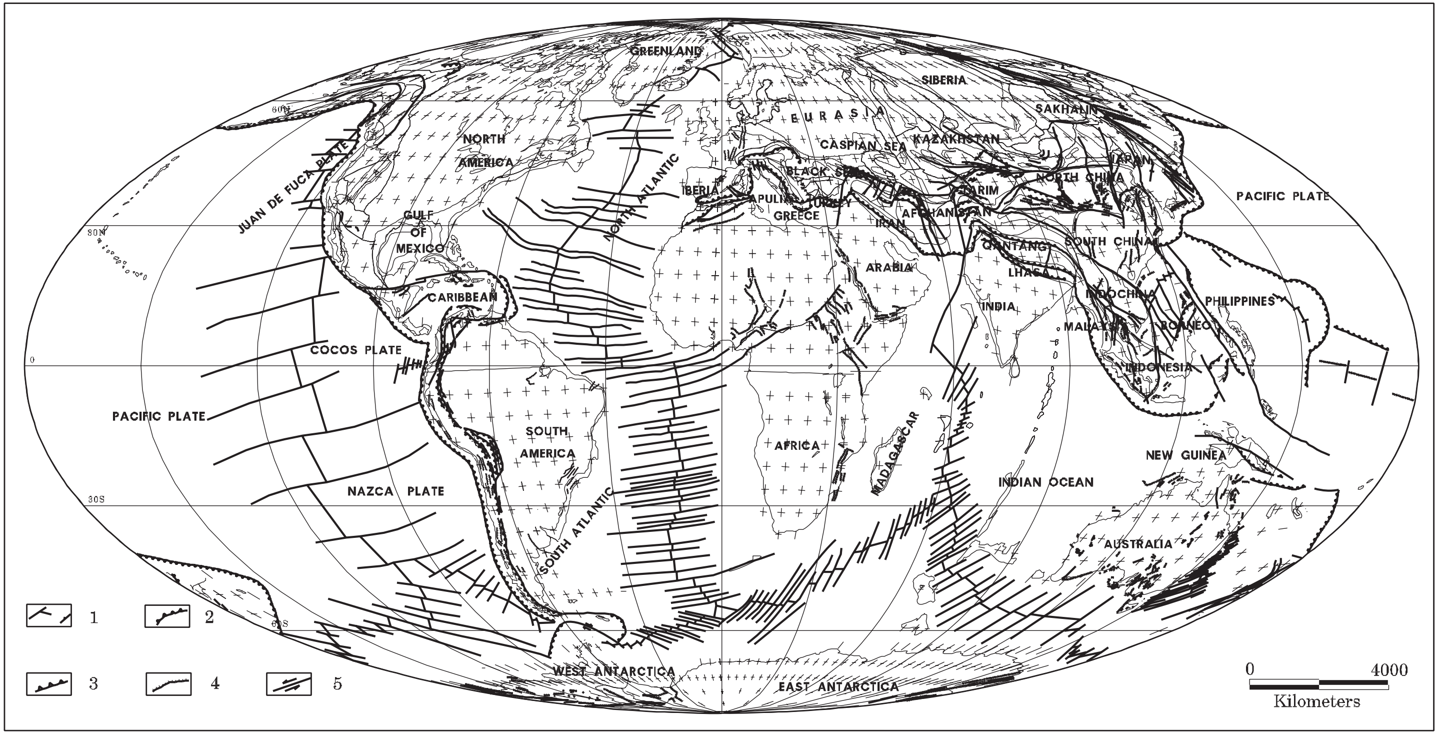

Fig. 21. Plate tectonic map of the Early Miocene (plates position as of 22 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 21. Mapa tektoniki płyt wczesnego miocenu (pozycja płyt 22 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

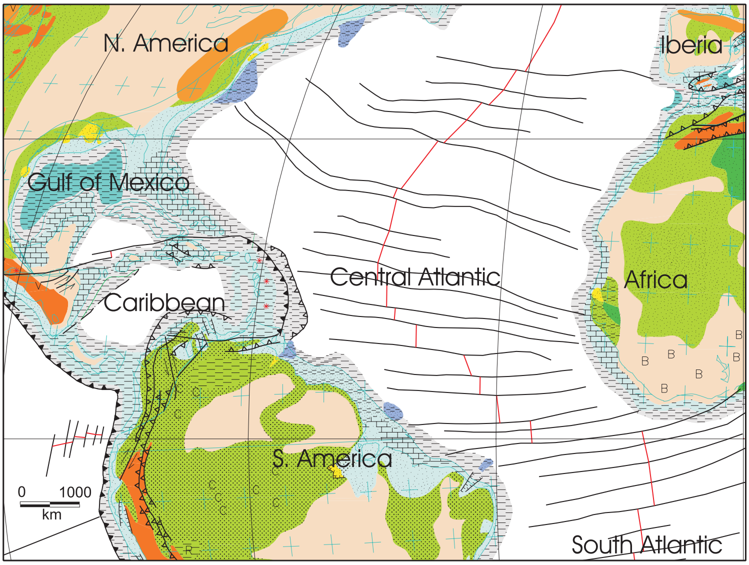

Fig. 22. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Early Miocene times

Fig. 22. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich we wczesnym miocenie

Both transpressive and transtensive deformations with rifts were present in this area. Fine-grained clastics dominated the entire Caribbean area. Volcanoes were active along Lesser Antilles (Donnelly 1973, Brown *et al*. 1977, Wadge 1984). The Amazonian realm as well as the area adjacent to the Andean thrustbelt in South America was covered by continental clastics with coals and with marine incursions in its eastern part (Fig. 22) (Pindell *et al*. 1991, Lamb *et al*. 1997, Olivero & Martinioni 2001, Golonka 2002, Ford & Golonka 2003, Hervouët *et al*. 2005, Jordan *et al*. 2007). Fine-grained clastics, locally with turbidites and carbonates, were deposited along South American eastern margin. Accretionary prism developed along the western margins of South America and Central America. Continental, deltaic and shallow-marine clastics, large carbonate platforms continued on Yucatan and on Bahama platform, as during previous time slices, and fine-grained clastics filled deeper part of the Gulf (Ronov *et al*. 1989, Galloway & Williams 1991, Brewster-Wingard *et al*. 1997, Xue 1997, Watkins 1999, Galloway *et al*. 2000, Budd 2002, Ford & Golonka 2003, Kiessling *et al*. 2004, Rosenfeld 2005, Wilson 2008).

The Atlas System in North Africa was uplifted and inverted supplying continental clastic to the surrounding areas (Fig. 22). The Alboran Sea extensional basin developed in the western Mediterranean behind the arc located between Iberia and Northern Africa (Morley 1993, Watts *et al*. 1993, Vissers *et al*. 1995). Fine-grained clastics and carbonates were deposited on the Northern African margins while interiors of Northern Africa like during the previous time slices were filled with continental clastics, mainly sands and gravel mounds (Said 1983, Coward & Ries 2003, Ford & Golonka 2003, Bumby & Guiraud 2005, Swezey 2009).

The collision of India and Eurasia continued. Metamorphism and crustal thickening reached a peak pre-20 Ma in eastern Kashmir (Searle 1996). The exhumation was widespread in the Himalayas (Yin & Harrison 2000, White *et al*. 2002, Steck 2003, Thiede *et al*. 2005, Kirstein *et al*. 2006, Caddick *et al*. 2007, Rutter *et al*. 2007). The development of the molasse basins continued in the Himalayan belt foreland (Burbank *et al*. 1996). Deltaic clastic sequences represented by fluviatile coastal plain, lagoonal and tidal flat complex, barrier island and offshore marine shelf were identified in several places along the eastern margin of India (Babu 2006). The large Ganges delta and fan was already developed during this time filled with deposits derived from the Himalayas (Najman *et al*. 2008.)

Rifting and formation of seaways with fine-grained clastic sedimentation continued in the Red Sea and in the Gulf of Aden while continental rifting continued in Ethiopia (Le Pichon & Francheteau 1978, White & McKenzie 1989, Huchon *et al*. 1991, Menzies *et al*. 1992, Vrielynck *et al*. 1997, Tesfaye *et al*. 2003, Hughes & Johnson 2005). The East Africa Rift System was in early stages of development (Girdler 1991, Golonka 2000, 2002, 2004, Golonka *et al*. 2006). Uplift and volcanics emplacement in the southern part of Arabian plate was related to these rifting events (Fig. 23). Carbonates were deposited on the north-eastern and eastern margins of Arabia (Whittle & Alsharhan 1994, Whittle *et al*. 1996, Alsharhan & Nairn 1997, Kiessling *et al*. 2003, Dull 2005, El-Saiy & Jordan 2007).

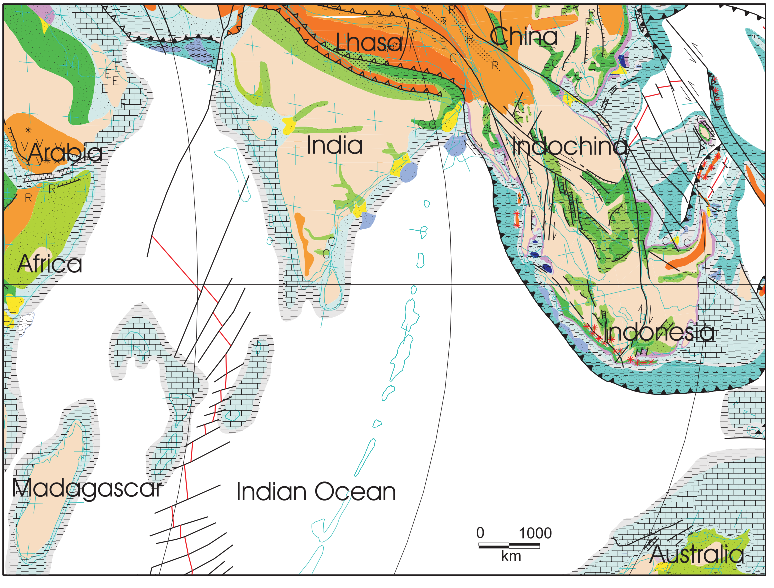

Fig. 23. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, Indian Ocean and adjacent areas during Early Miocene times

Fig. 23. Mapa tektoniki płyt, paleośrodowiska i litofacji Azji południowo-wschodniej, Oceanu Indyjskiego oraz obszarów sąsiednich we wczesnym miocenie

The Indo-Chinese pull-apart basins continued to develop. Subsidence was rejuvenated in the East China Sea (Kong 1998). The opening of the South China Sea continued, with a change in the direction from north-south to NW-SE (Lee & Lawver 1994). Sea-floor spreading began in the Japanese Sea. This back-arc spreading in the Sea of Japan was accompanied by rotation and movement of the Japanese plates (Ingle 1992, Tamaki *et al*. 1992, Jolivet *et al*. 1994, Kong 1998). Closing of the proto-South China Sea continued (Lee & Lawver 1994). The Australian plate continued a rapid northward movement, until the onset of the collision of its northern New Guinea margin with the Melanesian arc slowed down this motion (Longley 1997). Large carbonate platform developed on the northern Australian shelf (Davies *et al*. 1989, Cook 1990, Li & Powell 2001, Kiessling *et al*. 2003, Moss *et al*. 2004, Wilson 2008).

In the Red River Fault Zone, which marks the boundary between the South China and Indochina blocks, main activity occurred in two phases: during sinistral ductile shear active in 27–16 Ma, followed by exhumation and uplift from a depth of 20–25 km, and as dextral, predominantly brittle shear active in Plio-Quaternary times (*cf*. Allen *et al*. 1984, Tapponnier *et al*. 1990, Lacassin *et al*. 1993, Leloup *et al*. 1995, Cuong & Zuchiewicz 2001 and references therein, Golonka *et al*. 2006). Wang *et al*. (1998) maintain that the sinistral shearing took place between 27 and 17 Ma. However, recent fission-track studies indicate that the main period of ductile deformation in the RRFZ was finished by 25 Ma or 26 Ma (Żelaźniewicz *et al*. 2005, Golonka *et al*. 2006, Anczkiewicz *et al*. 2007), and new structural and geochronological data appear to document the polyphase ductile shear active between Early Cretaceous through Miocene times, which included dextral, sinistral, dextral transpression, and sinistral transtension regimes (Żelaźniewicz *et al*. 2005). The Late Miocene change of the sense of motion is commonly related to the history of collision between India and Eurasia (Tapponnier *et al*. 1986, 1990, Schaerer *et al*. 1994, Harrison *et al*. 1995, Golonka *et al*. 2006). The amount of sinistral offset along the Red River Fault Zone has been estimated at 330±60 km (Lacassin *et al*. 1993) to 500–700 km (Tapponnier *et al*. 1990, Leloup *et al*. 1995). This left-lateral transtension along The Red River Fault Zone caused rapid subsidence of the Song Hong Basin (Nielsen *et al*. 1999). The depositional environments in this basin varied widely during Late Oligocene-Early Miocene times, from fluvial, estuarine, and deltaic to offshore marine with the deposition of sandstones and mudstones. According to Hall *et al*. (2008), great thicknesses of Cenozoic sediments are present in Borneo and circum-Borneo basins. Carbonates were deposited on the Luconia platform (Zampetti *et al*. 2004, Bracco Gartner *et al*. 2005, Vahrenkamp *et al*. 2005).

A carbonate platform still persisted in the former Tethyan region in North Africa, the Taurus-Zagros area and in small areas in Turkey and Greece (Figs 23, 24) (Dercourt *et al*. 1993, Philip *et al*. 1996, Kiessling *et al*. 2003, Scheibner & Speijer 2008, Wilson 2008).

Collisions continued in the area between Africa and Eurasia. Thrusting occurred in the Riff area in Africa and the Betic area in the southern Spain, due to the collision of the Alboran Sea arc (Morley 1993, Vissers *et al*. 1995). The movement of Corsica and Sardinia caused the plates to push eastwards in the future, resulting in deformation of the Alpine-Carpathian system (Golonka *et al*. 2000). This deformation reached as far as to Romania and continued throughout the Neogene. The Calabrian terranes in the Western Mediterranean continued to progress eastwards (Dewey *et al*. 1989, van Dijk & Okkes 1991).

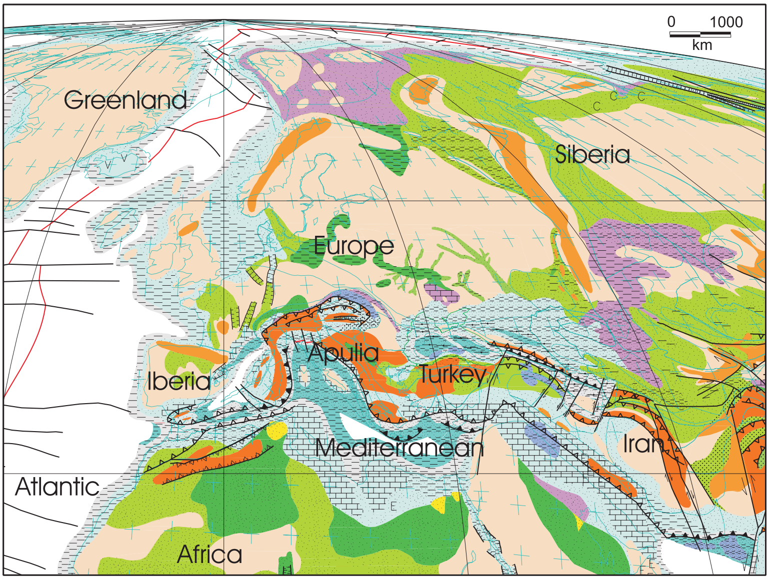

Fig. 24. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Early Miocene times

Fig. 24. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich we wczesnym miocenie

The thrust-and-foldbelt of the Apennines began to develop (Pialli & Alvarez 1997). The Apulia and the Alpine-Carpathian terranes were moving northwards, colliding with the European plate, until 17 Ma (Decker & Peresson 1996). Oblique collision between the North European plate and the overriding Western Carpathian terranes led to the development of the outer accretionary wedge, the built up many flysch nappes and the formation of a foredeep (Golonka *et al*. 2000). This process was completed in the Vienna basin area and then progressed northeastwards (Oszczypko 1997, 1998). The Eastern Mediterranean Sea (Fig. 24) began to be subducted beneath the newly formed Eurasian margin (Vrielynck *et al*. 1997). The subduction active zone was located north of the Ionian and Levantine basin. Crustal extension of the internal zone of the Alps started in the Early Miocene, during the continued thrusting (Decker & Peresson 1996). The Early to Middle Miocene extension and back-arc type rifting resulted in the formation of the intramountain Pannonian basin (Fig. 24) in Central Europe (Royden 1988, Decker & Peresson 1996, Tari *et al*. 1997, Golonka *et al*. 2000, Golonka 2004, 2006). According to Hámor *et al*. (2001), 24–18 Ma Pannonian Basin entered its early synrift phase. A new period of extension began in the Pontides-Sakariya continent in Turkey (Yilmaz *et al*. 1997, Golonka 2004). The collision of India and Eurasia influenced the Central Asia Area through the development of far-reaching strike-slip faults. Several blocks were deformed and thrust over the Turan platform in the Pamir, Afghan-Tadjik and Gissar areas. The Miocene phase of thrusting and folding of the Kopet-Dagh Mountains in Central Asia, with a strong strike-slip component was a result of the final stage of collision of the Lut plate with Eurasia. The Greater Caucasus Ocean was closed as a result of the collision of the Lesser Caucasus and Transcaucasus blocks with the Scythian platform, and the Caucasus Mountains began to form (Zonenshain *et al*. 1990, Kazmin 1991, Kopp 1997, Golonka 2004, 2007). The Paratethys continued its existence in Eastern Europe and Central Asia with mixed coarse and fine clastic deposition and with the Maykop source-rocks (Dercourt *et al*. 1993, Popov *et al*. 1993, 2006, Abrams & Narimanov 1997, Inan *et al*. 1997, Devlin *et al*. 1999, Rögl 1999, Devlin 2000, Golonka *et al*. 2000, 2006, Il'ina 2000, Brunet *et al*. 2003, Nevesskaya *et al*. 2003, Oszczypko 2006, Golonka 2007a, Piller *et al*. 2007, Stolyarov & Ivleva 2007, Hudson *et al*. 2008). In front of the Carpathian nappes, remnant flysch basin turned into molasse basin (Golonka *et al*. 2000, 2006). Limnic environment with fine-grained clastic and with first Miocene lignites developed in the Central European lowlands (Ziegler 1982, 1988, 1890, Piwocki 1998, Reichenbacher 2000, Szynkiewicz 2000, Eissmann 2002, Ford & Golonka 2003, Rasser *et al*. 2008). Rifting continued in Western Europe (Ziegler 1988, 1990, 1992, Wilson & Downes 1991, Bois 1993, Wilson 1994, Schäfer *et al*. 2005). Fine-grained clastics and sands continued in the orth Sea, but coarse-grained braided fluvial systems developed south of Scandinavia due to the uplift (Berstad & Dypvik 1982, Ziegler 1988, 1990, 1992, Jordt *et al.* 1995, 2000, Eidvin & Rundberg 2007, Japsen *et al*. 2008, Rasmussen *et al*. 2008, Marcussen *et al*. 2009). The Atlantic margin in Norway was uplifted during the Neogene time (Jensen & Schmidt 1993). Intracontinental deformation in Eurasia led to the uplift and formation of the modern Ural Mountains (Puchkov 1997). Continental clastics were deposited in West Siberian, Kara Sea and Pechora basin (Fig. 24) (Meyerhoff 1983, Green *et al*. 1984, Ronov *et al*. 1989, Volkova & Kul'kova 1996, Kulkova & Volkova 1997, Tull 1997, Bogdanov *et al*. 1998, Musatov & Pogrebitskij 2000, Akhmet'ev *et al*. 2001, Golonka *et al*. 2003b, Kuz'mina & Volkova 2008). Marginal marine clastics and local uplifts occurred within the Barents Sea area (Rønnevik *et al*. 1982, Vorren *et al*. 1990, 1991, Skagen 1993, Johansen *et al*. 1993, Musatov & Pogrebitskij 2000, Shipilov *et al*. 2006, Rasmussen *et al*. 2008).

Spreading in the North Atlantic and Arctic Eurasian Basin continued, and Iceland formed as a volcanic platform astride the North Atlantic spreading ridge (Lawver & Müller 1994, Golonka 2000, 2002, Golonka & Bocharova 2000, Ford & Golonka 2003, Golonka *et al*. 2003b). Connection between the Atlantic and Eurasian Basin was well established during Early Miocene times (Fig. 24). Clastics, mainly fine-grained, locally with volcanics were deposited along North Atlantic and Eurasian margins (Ziegler 1988, 1990, Joy 1992, Bull & Masson 1996, Ford & Golonka 2003, Golonka *et al*. 2003b, Laberg *et al*. 2005, Agterberg *et al*. 2007, Rasmussen *et al*. 2008).

Spreading between Africa, South America and Antarctica as well as between Australia and Antarctica and Africa continued (Lawver *et al*. 1985, Nürnberg & Müller 1991, Lawver & Gahagan 1993, Golonka 2000, 2002, Golonka & Ford 2003, Macdonald *et al*. 2003). Right lateral strike-slip fault was active between Scotia Sea plate and Antarctica (Lawver *et al*. 1992, Maldonado *et al*. 1998, 2005, Golonka 2000, 2002, Barker 2001, Eagles *et al*. 2005). Mixed clastics were deposited on the South American, African and Antarctic margins (Fig. 25) with addition of carbonates in the South African area (Dingle & Hendry 1984, 1989, Clemson *et al*. 1997, Billups *et al*. 2002, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Maldonado *et al*. 2005, Wigley &

Compton 2006, Parras *et al*. 2008). Sedimentation of continental clastics continued within the African basins, also large areas of southern South America werecovered with continental clastics containing coals and volcanics in the Andean foreland (Pindell & Tabbutt 1995, Barreda 1996, Limarino *et al*. 2001, Olivero & Martinioni 2001, Rodriguez & Littke 2001, Ford & Golonka 2003, Dávila *et al*. 2004, Bumby & Guiraud 2005, Marenssi *et al*. 2005, Dávila & Astini 2007, Torres Carbonell *et al*. 2008).

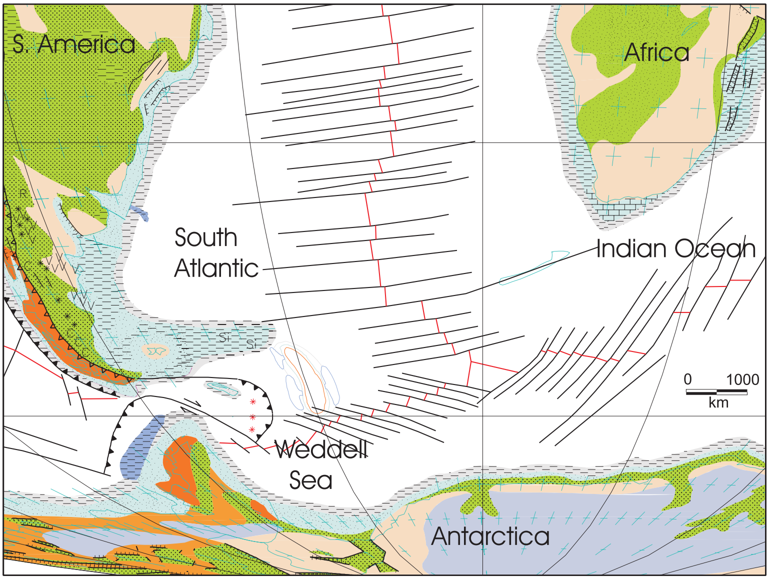

Fig. 25. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Early Miocene times

Fig. 25. Mapa tektoniki płyt, paleośrodowiska i litofacji Południowego Atlantyku, południowo-zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich we wczesnym miocenie.

Middle Miocene

Spreading in the Atlantic and Indian oceans continued, with a westward drift of the Americas and northward drifting of Africa, Eurasia and Australia (Fig. 26). South America moved faster than North America while Australia moved faster than Eurasia (Müller *et al*. 1997, Nürnberg & Müller 1991, Royer & Sandwell 1989). The rate of motion of South America again increased relative to the mantle (Pindell & Tabutt 1995).

The Andes were rejuvenated with crustal shortening, uplift and an increase of volcanic activity and with further development of foredeep filled with marine in the northern part, marginal marine and continental clastics, locally with coals (Pindell *et al*. 1991, Lamb *et al*. 1997, Olivero & Martinioni 2001, Golonka 2002, Ford & Golonka 2003, Hervouët *et al*. 2005, Duerto *et al*. 2006, Jordan *et al*. 2007). Continental clastics were also deposited within the Amazonian basins (Fig. 27).

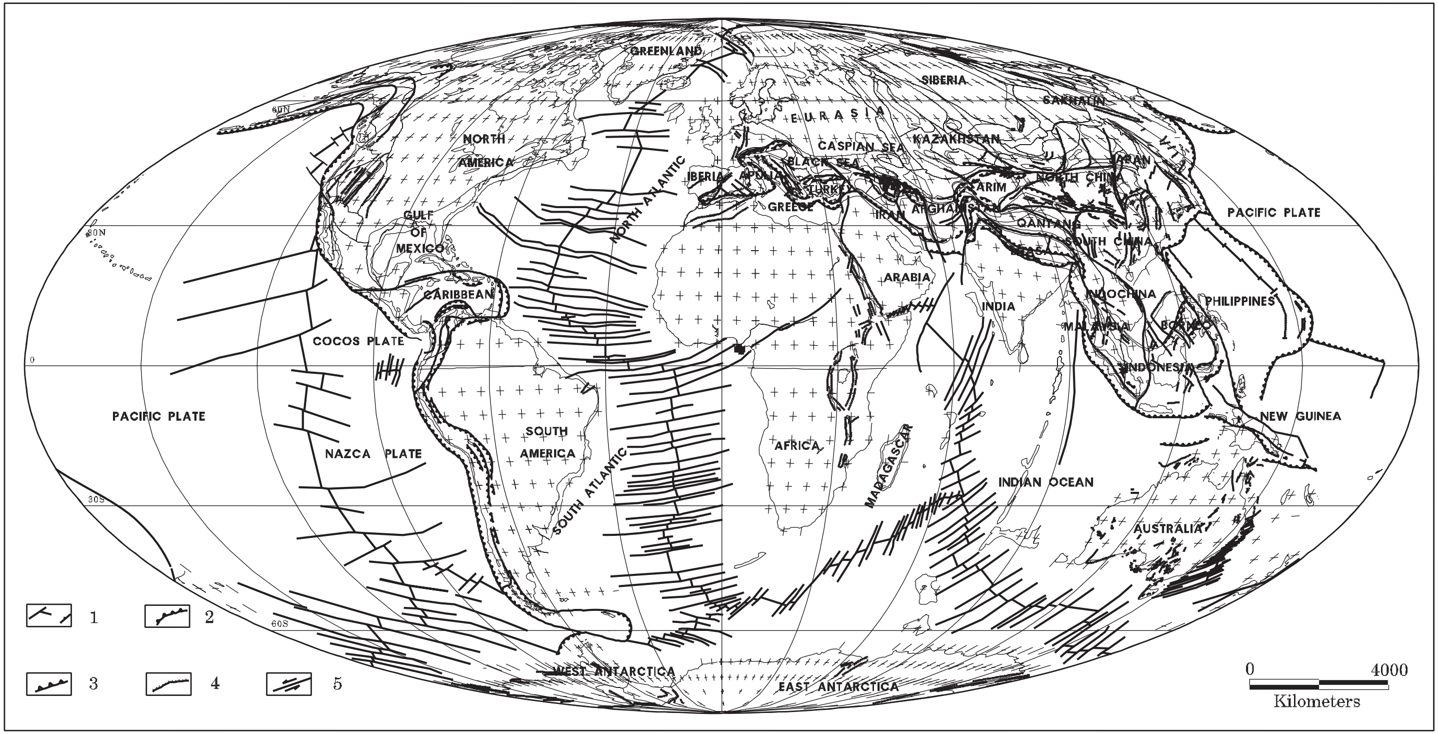

Fig. 26. Plate tectonic map of the Middle Miocene (plates position as of 14 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 26. Mapa tektoniki płyt środkowego miocenu (pozycja płyt 14 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

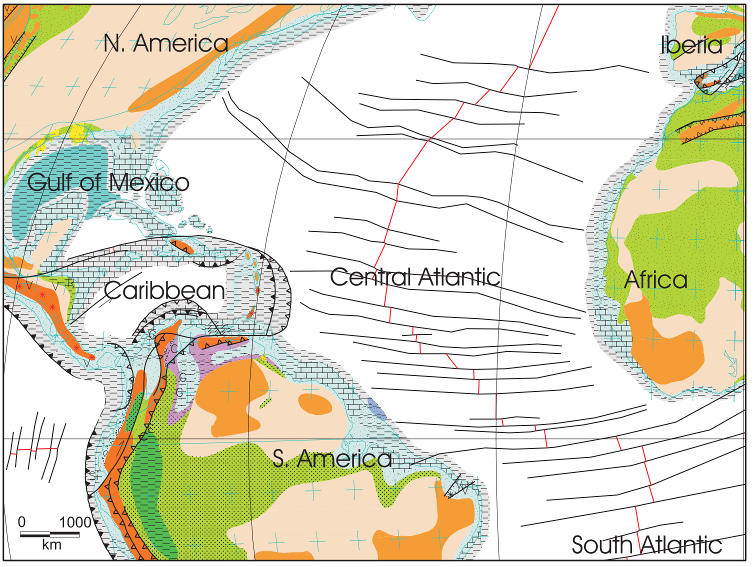

Fig. 27. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Middle Miocene times

Fig. 27. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich w środowym miocenie

Subduction zone developed along the southern margin of the Caribbean plate, while left-lateral strike-slip fault was still active along its northern margin (Burke 1988, Ross & Scotese 1988, Pindell & Barrett 1990, Hoorn *et al*. 1995, Ave Lallemant 1997, Golonka 2000, 2002, Audemard & Audemard 2002, Summa *et al*. 2003, Golonka *et al*. 2006, Iturralde-Vinent 2006, Pindell *et al*. 2006). Hispaniola moved to the position southwest of Cuba. Carbonates were abundant along the northern and southern Caribbean boundary and large carbonate platform existed between Nicaragua and Hispaniola, on Cuba, Yucatan, Florida and Bahamas (Brewster-Wingard *et al*. 1997, Budd 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Baron-Szabo 2005, Mutti *et al*. 2005, Rosenfeld 2005, Johnson & Pérez 2006, Wilson 2008). Fine-grained clastic were deposited in the Lesser Antilles back-arc basin and in the accretionary prism (Speed *et al*. 1989, Torrini & Speed 1989, Bouysse & Westercamp 1990, Donovan *et al*. 2003, Ford & Golonka 2003, Gorney *et al*. 2007). Deep Gulf of Mexico was still filled with deep-water clastics, with deltaic systems developed along the northern margin (Watkins 1999, Galloway *et al*. 2000). Deposition of fine-grained clastics, locally with turbidites and carbonates continued along South American eastern margin and Northern African western margins.

Continental clastics, mainly sands and gravel mounds filled basins within the interiors of Northern Africa like during the previous time slices (Said 1983, Coward & Ries 2003, Ford & Golonka 2003, Bumby & Guiraud 2005, Swezey 2009). The continued of thrusting occurred in the Riff area in Africa and in the Betic area in southern Spain as a result of the collision of the Alboran Sea arc with the Africa and Iberian plates (Morley 1993, Vissers *et al*. 1995).

The collision of India and Eurasia continued, with a strong, northwest motion component. Thrusting was active in the High Himalayas about 21–18 Ma (Fig. 28). The northward movement of India caused the opening of the Andaman Sea (Curray *et al*. 1982, Lee & Lawver 1994).

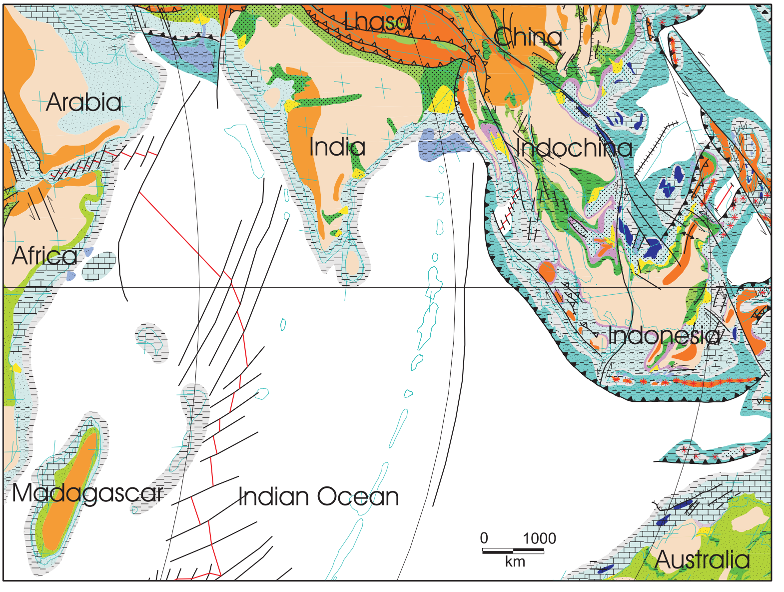

Fig. 28. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, Indian Ocean and adjacent areas during Middle Miocene times

Fig. 28. Mapa tektoniki płyt, paleośrodowiska i litofacji Azji południowo-wschodniej, Oceanu Indyjskiego oraz obszarów sąsiednich w środowym miocenie

The opening of the South China Sea ended (Fig. 28) (Taylor & Hayes 1980, Holloway 1982, Lee & Lawver 1994, Hall 1998, 2002). The Sulu Basin in SE Asia opened as a result of the subduction along the northeastern margin of Kalimantan (Lee & Lawver 1994). Spreading in Marianas, in the Eastern Pacific, began at 15 Ma (Scott & Kroenke 1980). The Sea of Japan went through its second episode of opening (Ingle 1992, Tamaki *et al*. 1992, Jolivet *et al*. 1994). In Southeast Asia, the North Palawan microcontinent collided with Kalimantan and with the West Philippine Archipelago (Taylor & Hayes 1980, 1983, Holloway 1982, Lee & Lawver 1994, Almasco *et al*. 2000). The collision with the Philippines resulted in a change of subduction polarity to a westward one, along the east margin of Philippines, at the Philippine trench (Lee & Lawver 1994). An eastward-dipping subduction zone developed along the Manila Trench. The left-lateral Philippine fault initiated motion between the East and West Philippines (Fig. 28). Motion began along the Sumatra Fault System at this time, and was responsible for crustal shortening in northwestern Sumatra (Huchon & LePichon 1984, Lee & Lawver 1994). Southwestern Japan rapidly rotated in a clockwise direction (Ingle 1992, Tamaki *et al*. 1992, Jolivet *et al*. 1994). The Izu-Bonin Arc collided with central Japan (Kong 1998, Kong *et al*. 2000).

As was mentioned above, the left-lateral movement along the Red River Fault Zone ceased by Mid-Miocene. The change of displacement direction is expressed in the Song Hong Basin by the distinct unconformity near the base of Mid-Miocene, which in places shows deep channel incision and lateral shift of depocenters (Nielsen *et al*. 1999). Thick prograding deltaic units of sandstones, siltstones, mudstones, and brown coal were deposited in Middle-Late Miocene times in the northern part of the basin as well as in the onshore Hanoi trough (Fig. 28). According to Hutchison (2004), the end of rifting in the southern part of South China Sea is expressed by Mid-Miocene unconformity followed by thick post-rift sequences. According to Wang *et al*. (2000), the cessation of sea-floor spreading in South China, coincide with continuous basaltic eruption since ~16 Ma. This igneous activity produced alkali basalts, basanite, trachy basalts and basaltic trachy andesite. They are widely distributed in Vietnam. Clastics prevailed in forarc basins along the Sunda Arc offshore Indonesia (Hutchison 1989, Golonka *et al*. 2006).

The Gulf of Aden went through the spreading phase (Fig. 28), while rifting continued in the Red Sea and East Africa Rift System (Le Pichon & Francheteau 1978, White & McKenzie 1989, Girdler 1991, Huchon *et al*. 1991, Menzies *et al*. 1992, Vrielynck *et al*. 1997, Golonka 2000, 2002, 2004, Tesfaye *et al*. 2003, Hughes & Johnson 2005, Golonka *et al*. 2006).

Mountains reached its main thrusting stage (Fig. 29) due to collision of Arabia and Iranian plates (Berberian & Berberian 1981, Dercourt *et al*. 1993, Alavi 1994, Golonka 2004, Fakhari *et al*. 2008, Heydari 2008). Foredeep developed in front of the Zagros thrust.

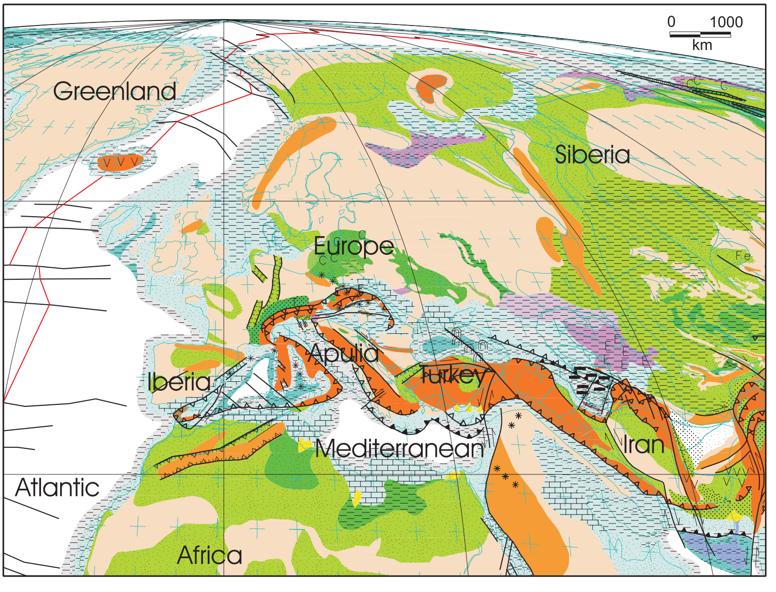

Fig. 29. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Middle Miocene times

Fig. 29. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich w środkowym miocenie

Southwestern part of Arabia was uplifted while clastic sedimentation prevailed in its northeastern part with remnant carbonate platform in the center (Whittle & Alsharhan 1994, Whittle *et al*. 1996, Alsharhan & Nairn 1997, Kiessling *et al*. 2003, Dull 2005, El-Saiy & Jordan 2007). Volcanics were emplaced in the northwestern part of this plate (Saif & Shah 1988, Lustrino & Sharkov 2006, Krienitz *et al*. 2007).

The collision of India and Eurasia influenced the Central Asia area through the development of far-reaching strike-slip faults (Fig. 29). Several blocks were deformed and thrust over the Turan platform in the Pamir, Afghan-Tadjik and Gissar areas. The Miocene phase of thrusting and folding of the Kopet-Dagh Mountains in Central Asia, with a strong strike-slip component was a result of the final stage of collision of the Lut plate with Eurasia. The Greater Caucasus Ocean was closed as a result of the collision of the Lesser Caucasus and Transcaucasus blocks with the Scythian platform, and the Caucasus Mountains began to form (Zonenshain *et al*. 1990, Kazmin 1991, Kopp 1997, Golonka 2000, 2004, 2007a). The Alborz trough in the South Caspian Sea opened, and extension progressed into the West Turkmen Depression in Central Asia (Golonka 2007a). The southwestern part of the South Caspian basin was reopening, while the northwestern part was gradually reduced in size. The South Caspian Microcontinent or Godin uplift (e.g. Nadirov *et al*. 1997) separated the southwestern part of the South Caspian basin and the Western Turkmenistan area.

The opening of the Tyrrhenian Sea (Spadini *et al*. 1995) as well as Valencia Trough (Vegas 1992, Torres *et al*. 1993, Golonka 2004) was initiated. Extension in the Alpine--Carpathian system continued. Extension also occurred in the Apennines (Carmignani *et al*. 1994, Anelli *et al*. 1996, Golonka 2004). Strike-slip in the Pannonian basin (Decker & Peresson 1996) contributed to the formation of pull-apart elements of the Pannonian system. Tertiary magmatism was crossing the Carpathians between Moravia and Upper Silesia, on one side, and the Pannonian Basin, on the other. Mantle doming contributed to crustal stretching (Golonka *et al*. 2000, Golonka & Bocharova 2000, Golonka 2004, Golonka & Picha 2006). The Calabrian arc and subduction zone collided with Africa and the Southern Sicilian-Maltese platform (Dewey *et al*. 1989, van Dijk & Okkes 1991, Ricou 1996). The wing of this collision formed the southern Apennines. Thrusting also continued in the northern Apennines. The formation of the West Carpathian thrusts was completed (Golonka *et al*. 2000, Golonka 2004). The thrust front was still migrating eastwards in the Eastern Carpathians. The Paratethys continued its existence in Eastern Europe and Central Asia with mainly fine clastic and carbonate deposition (Fig. 29) (Dercourt *et al*. 1993, Rögl 1999, Golonka *et al*. 2000, 2006, Il'ina 2000, Nevesskaya *et al*. 2003, Oszczypko 2006, Popov *et al*. 2006, Piller *et al*. 2007). Brown coals were abundant in the limnic environments in the Central European lowlands (Ziegler 1982, 1988, 1890, Piwocki 1998, Reichenbacher 2000, Szynkiewicz 2000, Eissmann 2002, Ford & Golonka 2003, Rasser *et al*. 2008). Rifting continued in Western Europe (Ziegler 1988, 1990, 1992, Bois 1993, Wilson & Downes 1991, Wilson 1994, Schäfer *et al*. 2005).

Spreading in the North Atlantic and Arctic Eurasian Basin continued (Figs 26, 29) and Iceland formed as a volcanic platform astride the North Atlantic spreading ridge (Lawver & Müller 1994). Strike-slip motion was initiated between Greenland and Svalbard. Fine--grained clastics were deposited along North Atlantic and Eurasian margins (Ziegler 1988, 1990, Joy 1992, Bull & Masson 1996, Ford & Golonka 2003, Golonka *et al*. 2003b, Laberg *et al*. 2005, Agterberg *et al*. 2007, Rasmussen *et al*. 2008).

Areas of continental clastic deposition in West Siberian, Kara Sea and Pechora basin (Fig. 28) were enlarged (Meyerhoff 1983, Green *et al*. 1984, Ronov *et al*. 1989, Volkova & Kul'kova 1996, Kulkova & Volkova 1997, Tull 1997, Bogdanov *et al*. 1998, Musatov & Pogrebitskij 2000, Akhmet'ev *et al*. 2001, Golonka *et al*. 2003b, Kuz'mina & Volkova 2008). Marginal marine and shallow-marine clastic deposition was also present in the south-eastern Kara Sea. Miocene marginal marine, shallow marine and continental clastics with lignites occurred in New Siberian Islands (Kos'ko & Trufanov 2002). Continental clastic deposition prevailed in the Barents Sea area, shallow marine and marginal environment occurred only in the southern part of the basin (Ronnevik *et al*. 1982, Vorren *et al*. 1990, 1991, Skagen 1993, Johansen *et al*. 1993, Musatov & Pogrebitskij 2000, Shipilov *et al*. 2006, Rasmussen *et al*. 2008). A seaway with fine-grained sedimentation connected Barents and Kara seas.

Spreading in the southern Atlantic oceans continued, with a westward drift of South America and northward drifting of Africa (Lawver *et al*. 1985, Nürnberg & Müller 1991, Lawver & Gahagan 1993, Golonka 2000, 2002, Golonka & Ford 2003, Macdonald *et al*. 2003). A right lateral strike-slip fault was still active between Scotia Sea plate and Antarctica during Middle Miocene times (Lawver *et al*. 1992, Maldonado *et al*. 1998, 2005, Golonka 2000, 2002, Barker 2001, Eagles *et al*. 2005). Mixed clastics were deposited on the South American, African and Antarctic Margins (Fig. 30) with addition of carbonates offshore southern Brasil (Dingle & Hendry 1984, 1989, Clemson *et al*. 1997, Billups *et al*. 2002, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Maldonado *et al*. 2005, Wigley & Compton 2006, Parras *et al*. 2008).

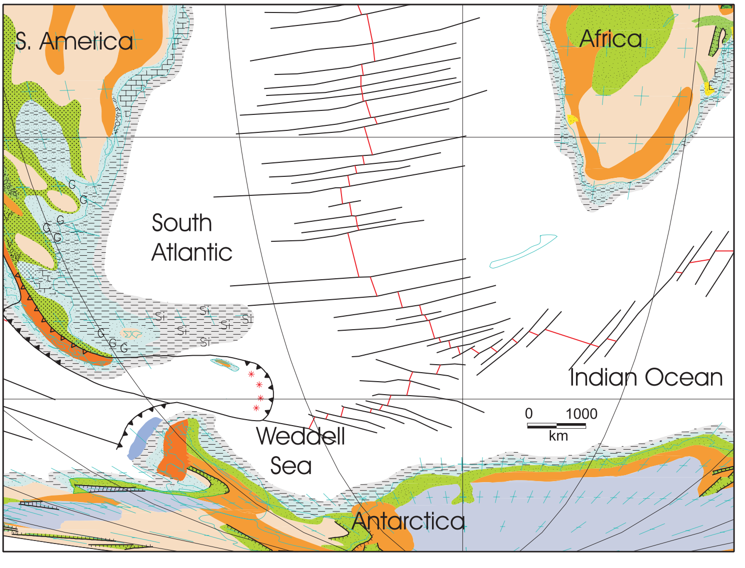

Fig. 30. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Middle Miocene times

Fig. 30. Mapa tektoniki płyt, paleośrodowiska i litofacji południowego Atlantyku, południowo-zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich w środowym miocenie

The rate of motion of South America again increased relative to the mantle (Pindell & Tubutt 1995). The Andes were rejuvenated with crustal shortening, uplift and an increase of volcanic activity. The marine inundation of the Argentina coastal basins occurred during the Neogene high sea-level and climatic optimum at 17–15 Ma. This optimum was followed by a cooling period. According to Lewis *et al*. (2007), the glacial record from Antarctica provides terrestrial evidence linking middle Miocene global climate cooling to a permanent reorganization of the Antarctic cryosphere and to subsequent growth of the polar East Antarctic Ice Sheet.

Sedimentation of continental clastics continued within the African basins and of part of southern South America in the Andean foreland (Pindell & Tabbutt 1995, Barreda 1996, Limarino *et al*. 2001, Olivero & Martinioni 2001, Rodriguez & Littke 2001, Ford & Golonka 2003, Dávila *et al*. 2004, Bumby & Guiraud 2005, Marenssi *et al*. 2005, Dávila & Astini 2007, Torres Carbonell *et al*. 2008). Deltas of Orange and Limpopo rivers supplied coarse clastics (Goudie 2005, Bluck *et al*. 2007). Clastics, mainly fine-grained with silica, diatomaceous ooze and glauconite were deposited on Falkland Plateau (Muza & Wise 1983, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003).

Late Miocene-Pliocene

This was the time of the assembly of continents (Fig. 31). Large masses of continents and continental shelves were situated around the North Pole. The mountain building process continued in the Andes (Pindell & Tabbutt 1995). Late Miocene-Pliocene compression occurred also in East Venezuela (Eva *et al*. 1989). Shortening took place in Central Andes (Lamb *et al*. 1997). Crustal shortening, nappe emplacement and uplift of Northern Andes were followed by further development of foredeeps (Fig. 32). These foredeeps were filled mainly with continental coarse-grained clastics (Pindell *et al*. 1991, Lamb *et al*. 1997, Olivero & Martinioni 2001, Golonka 2002, Ford & Golonka 2003, Hervouët *et al*. 2005, Duerto *et al*. 2006, Jordan *et al*. 2007). Marine incursions occurred only in Venezuela. Also inter-Andean basins had developed (Pindell & Tabbutt 1995). Erosion of Andes produced massive volume of molasse detritus. The Amazonian basin became narrow, with coarse continental clastic deposition. Increased orogenic activity enhanced continental drainage Narrow continental margins with deltaic deposits were common along the north-eastern coast of South America. Large volumes of clastic sediments were delivered to deltas and the deep-sea Amazon and Orinoco fans (Fig. 32). Increased deposition on continental shelves helped to bury and mature older organic-rich deposits. Carbonates played minor role in this area.

A strike-slip fault along the northern margin of Caribbean area displayed only minor activity, while subduction zone was active along the southern margin of the Caribbean plate (Burke 1988, Ross & Scotese 1988, Pindell & Barrett 1990, Hoorn *et al*. 1995, Ave Lallemant 1997, Golonka 2000, 2002, Audemard & Audemard 2002, Summa *et al*. 2003, Golonka *et al*. 2006, Iturralde-Vinent 2006, Pindell *et al*. 2006). On the Atlantic side, subduction was active along the southern part of the Lesser Antilles arc (Fig. 32). Deposition of fine-grained clastics continued the Caribbean-Atlantic margins as well as in the Lesser Antilles back-arc basin (Speed *et al*. 1989, Torrini & Speed 1989, Bouysse & Westercamp 1990, Donovan *et al*. 2003, Ford & Golonka 2003, Gorney *et al*. 2007).

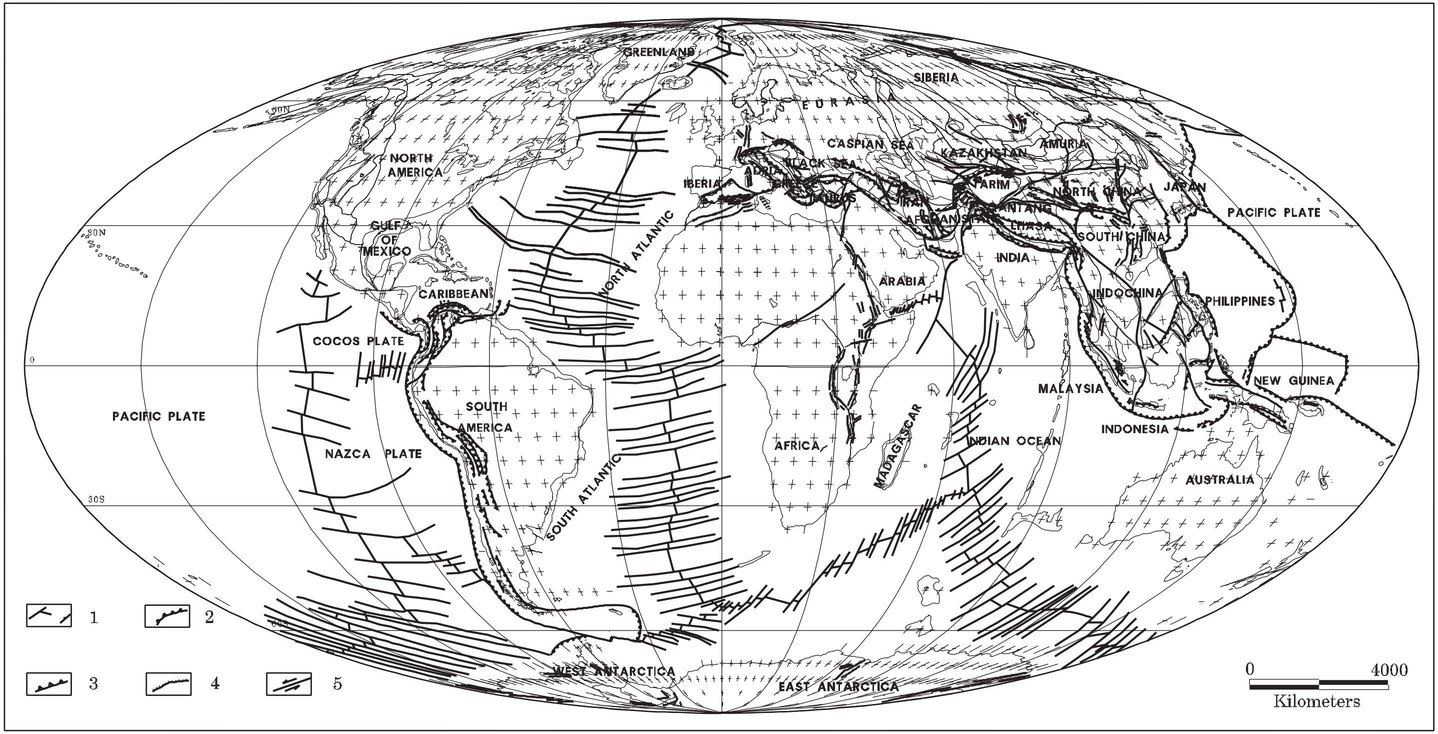

Fig. 31. Plate tectonic map of the Late Miocene-Pliocene (plates position as of 6 Ma). Modified from Golonka (2002): 1 – oceanic spreading center and transform faults, 2 – subduction zone, 3 – thrust fault, 4 – normal fault, 5 – transform fault

Fig. 31. Mapa tektoniki płyt późnego miocenu-pliocenu (pozycja płyt 6 milionów lat temu). Zmienione wg Golonki (2002): 1 – centrum spredingu oceanicznego i uskok transformujący, 2 – strefa subdukcji, 3 – nasunięcie, 4 – uskok normalny, 5 – uskok przesuwczy

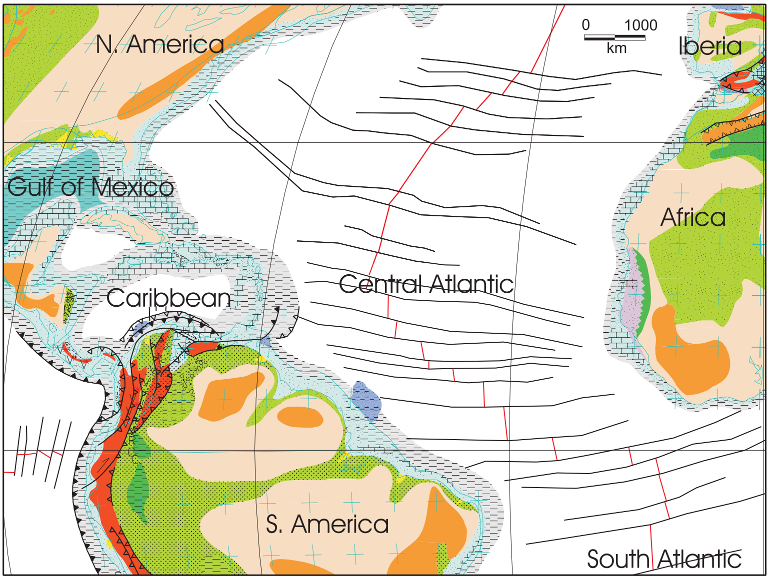

Fig. 32. Plate tectonic, paleoenvironment and lithofacies map of Central Atlantic and adjacent areas during Late Miocene-Pliocene times

Fig. 32. Mapa tektoniki płyt, paleośrodowiska i litofacji Atlantyku centralnego oraz obszarów sąsiednich w późnym miocenie-pliocenie

Deposition of carbonates continued on still active large platforms on Yucatan, southern part of Florida and adjacent Gulf of Mexico, and between Nicaragua and Hispaniola, while most of Cuba was uplifted and emerged (Brewster-Wingard *et al*. 1997, Budd 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Baron-Szabo 2005, Mutti *et al*. 2005, Rosenfeld 2005, Johnson & Pérez 2006, Wilson 2008).

The sea-way in the Central American isthmus existed during Late Miocene and was finally closed during the Pliocene (e.g. Iturralde-Vinent 2006). According to Coates *et al.* (2004), no Pliocene deposits are recorded from either the Darien or the Panama Canal Basin, and no sediments younger than 4.8 Ma have been identified in the Atrato Basin of Colombia, suggesting rapid uplift and extensive emergence of the isthmus in the latest Miocene. The final closure of the Central American Seaway (e.g. Bartoli *et al.* 2005) induced an increased poleward salt and heat transport, strengthening of North Atlantic, intensification of moisture supply to northern high latitudes. The glaciation in the Northern Hemisphere followed the great, irreversible "climate crash" at marine isotope stage 2.74 Ma (Bartoli *et al*. 2005), which was closely related to closure of Central American seaways. According to Nesbitt and Young (1997), dramatically increased pelagic sedimentation rates in the Venezuelan Basin since 4 Ma suggest intensification of the Atlantic equatorial current and strengthening of equatorial Atlantic summer storms (hurricanes), both of which strengthened the Gulf Stream current.

According to Frizon de Lamotte *et al*. (2009 in press), the Late Miocene-Pliocene was time of the inversion tectonics, crustal thickening and moderate uplift of the Atlas

Mountains in Africa (Figs 32, 34). This event was related to the general uplift of continent and narrowing of West Africa-Atlantic margins shelves (Golonka 2000, 2002, Ford & Golonka 2003). Carbonate sedimentation was extensive along the Western African margins (Ford & Golonka 2003, Kiessling *et al*. 2003) According to Wagner (2002), the Latest Miocene-Early Pliocene organic carbon deposition was closely linked to the evolution of the African trade winds, continental upwelling in the eastern Equatorial Atlantic, ocean chemistry and eustatic sea level fluctuations.

Many deltas developed along the Indian Ocean (Fig. 33) margins (Agarwal *et al*. 1996, Qayyum *et al*. 1997, Uddin & Lundberg 1998, 2004, Behera *et al*. 2004, Golonka *et al*. 2006, Bouillon *et al*. 2007, Robinson *et al*. 2007, Coleman *et al*. 2008). Indus and Ganges deltas and fans were supplied by massive input of clastic material from erosion of uplifted Himalayas (Giosan *et al*. 2006). Process of mountain building in the Himalayas and Central Asia continued. Metamorphism and crustal thickening reached a peak about 11–4 Ma, in Nanga Parbat in the High Himalayas (Searle 1996). Uplift, exhumation and formation of topographic highs followed. The development of the molasse basins continued in the Himalayan belt foreland (Burbank *et al*. 1996). Terrestrial, mainly coarse-grained clastics were deposited there. Like during the previous time slice India's interior was uplifted, locally with continental clastics. The opening of the Andaman Sea continued through Late Miocene and Pliocene times to the present day (Lee & Lawver 1994, Golonka *et al*. 2006). Subduction along the Manila Trench was caused the collision of the North Luzon Arch with the East Asian continental margin at Taiwan (Suppe 1981, Teng 1990, Lee & Lawver 1994, Golonka *et al*. 2006).

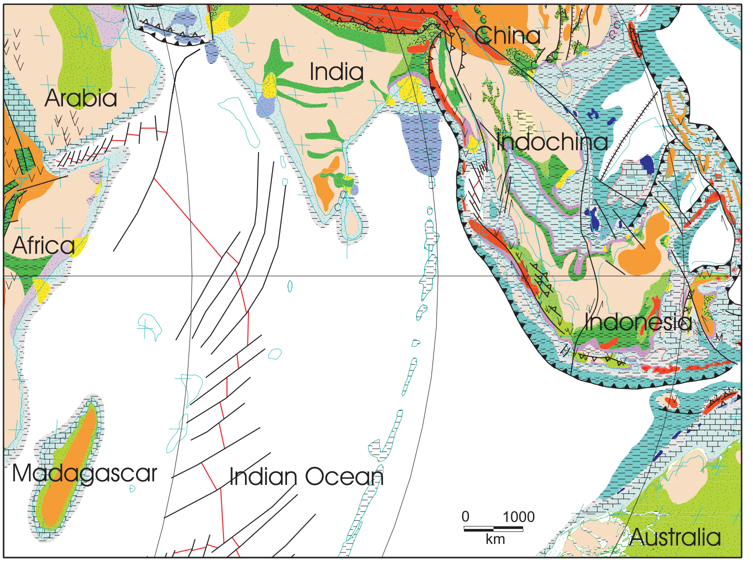

Fig. 33. Plate tectonic, paleoenvironment and lithofacies map of southeastern Asia, Indian Ocean and adjacent areas during Late Miocene-Pliocene times

Fig. 33. Mapa tektoniki płyt, paleośrodowiska i litofacji Azji południowo-wschodniej, Oceanu Indyjskiego oraz obszarów sąsiednich w późnym miocenie-pliocenie

The Taiwanese thrustbelt was formed during this second phase of collision, with an estimated 200–300 km of crustal shortening. Northward movement of Australia continued with compression in the New Guinea-Melanesian Arc. The Timor area of the Australian continent collided with the Sunda Arc (Longley 1991, Golonka *et al*. 2006).

The Red River Fault Zone in Vietnam and adjacent part of South China went into predominantly brittle shear active phase (*cf*. Allen *et al*. 1984 Tapponnier *et al*. 1990 Lacassin *et al*. 1993, Leloup *et al*. 1995, Cuong & Zuchiewicz 2001 and references therein, Golonka *et al*. 2006). In the Song Hong Basin, according to Nielsen *et al*. (1999), the strike-slip activity caused inversion structures truncated by the Late Miocene unconformity. Renewed and increased subsidence of the basin resulted in thick Latest Miocene to Quaternary section overlying the unconformity. South of Hainan, the thickness of Pliocene shelfal to bathyal mudstones is up to 5 km (Nielsen *et al*. 1999). The deposits in the northern part of the basin are dominated by offshore marine to shallow-marine mudstones, siltstones, and sandstones, with minor proportions of lagoonal and possibly fluvial deposits (Nielsen *et al*. 1999, Golonka *et al*. 2006). Deposition of carbonates on the Luconia platform continued (Zampetti *et al*. 2004, Bracco Gartner *et al*. 2005, Vahrenkamp *et al*. 2005). A large carbonate platform existed on the northern Australian shelf (Davies *et al*. 1989, Cook 1990, Li & Powell 2001, Kiessling *et al*. 2003, Moss *et al*. 2004, Wilson 2008).The spreading in the Gulf of Aden became a part of the main spreading of the Indian Ocean (Royer & Sandwell 1989). Arabia and Ethiopia were uplifted with emplacement of alkali basaltic volcanics on both sides of Red Sea (Le Pichon & Francheteau 1978, White & McKenzie 1989, Huchon *et al*. 1991, Menzies *et al*. 1992, Vrielynck *et al*. 1997, Tesfaye *et al*. 2003, Hughes & Johnson 2005). According to Bosworth *et al*. (2005), in the Early Pliocene the influx of marine waters through Bab al Mandeb increased and Red Sea sedimentation thereafter returned to predominantly open marine conditions with deposition of carbonate muds and reef facies (Figs 33, 34). Accretionary prism with large turbiditic fans developed along the active margin Arabian Sea in South Iran and Pakistan (Platt *et al*. 1985, Critelli *et al*. 1990, McCall 2002, Golonka 2004, Golonka *et al*. 2006). The uplifted Zagros Mountains supplied huge amount of continental molasse coarse-clastic deposits to their foredeep (Berberian & Berberian 1981, Dercourt *et al*. 1993, Alavi 1994, Golonka 2004, Fakhari *et al*. 2008, Heydari 2008). Fine-grained to pebbly coarse-grained fluvial sandstones of the Late Miocene to Pliocene were deposited in eastern Saudi Arabia (Nasir *et al*. 2007).

Collision of Indian continent and Lut plate with Eurasia caused deformation of the Central Asia region (Fig. 34). The system of NW-SE transform faults was developed. These faults were a predominant plate tectonic force in the Turan platform, Kopet Dagh area (Trifonov 1978, Kopp 1997, Lyberis & Manby 1999b, Golonka 2004, 2007b), and strongly influenced the South Caspian region. The deformation connected with the SE-NW strike slip faults were observed in Great Balkhan Area, Apsheron ridge, South Caspian area, Alborz Mountains, and Kura Basin. The N-S strike slip movement system was probably still active, but dramatically reduced. The subduction zone south of the Apsheron ridge became passive perhaps at the end of the Miocene, because of this SE-NW movement of the lithospheric plates. The collision between the South Caspian Microcontinent and Scythian-Turan plate was never concluded. It appears, however, that the Apsheron subduction zone is active again today (Artemjev & Kaban 1994, Priestley *et al*. 1994, Golonka 2004, 2007b). The Jurassic-Cretaceous back-arc system oceanic and attenuated crust in the Cheleken and South West Caspian basin as well as Tertiary oceanic and attenuated crust in the Alborz basin and part of the South West Caspian Basin were locked between adjacent continental plates and orogenic systems.

Maximum subsidence of the South Caspian Basin took place mainly during the Pliocene, when more then 8,000 m to 10,000 m of sediment known as Productive Series and Variegated Series were deposited. The isolation of the Paratethys-Caspian Sea caused changes in the water salinity. Generally, the sediments in the South Caspian area, as well as in the adjacent basin on the Scythian-Turan platform, in West Turkmenia and Kura Basin were deposited in the marginal marine environment. The Paleo-Volga, Paleo-Amu-Daria and Paleo-Kura rivers delivered majority of sediments.

Continuation of thrusting occurred in the Riff area in Africa as well as the Betic area in southern Spain (Fig. 34) due to the collision of the Alboran Sea arc (Morley 1993). This thrusting temporarily cut off the Mediterranean Sea from the Atlantic, in the Gibraltar Strait area, causing the Messinian salinity crisis (Hsü *et al*. 1973, Ruggieri & Sprovieri 1976, Müller & Mueller 1991, Rouchy & Saint Martin 1992, Blanc 2006, Rouchy & Caruso 2006, Ryan 2008). Deposition of the 3-km-thick evaporite suite of carbonates, marls, sulfates, halite, potash as well as debris from the basin margin erosion took place in Mediterranean Sea during this crisis (Fig. 34). Compressional thrusting continued in the Calabrian Arc, with accompanying strike-slip faulting and change of rotation to SE (Dewey *et al*. 1989, van Dijk & Okkes 1991, Golonka 2000, 2002, 2004).

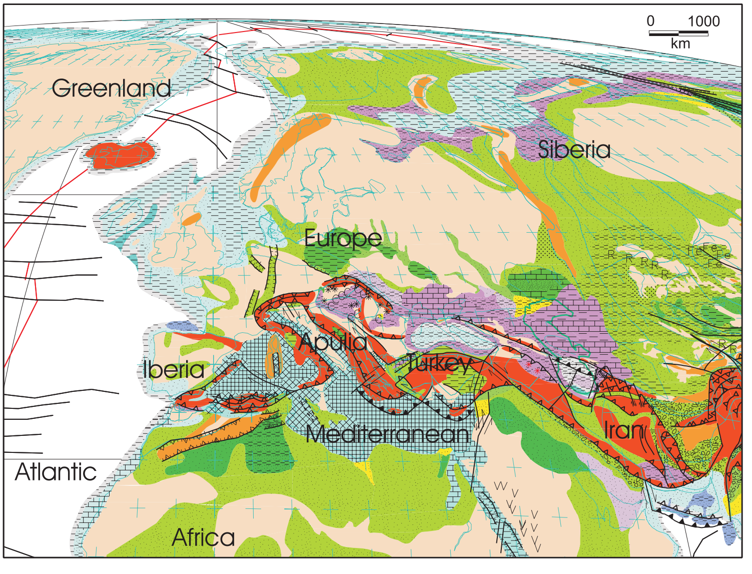

Fig. 34. Plate tectonic, paleoenvironment and lithofacies map of Europe and adjacent areas during Late Miocene-Pliocene times

Fig. 34. Mapa tektoniki płyt, paleośrodowiska i litofacji Europy oraz obszarów sąsiednich w późnym miocenie-pliocenie

The Tyrrhenian Sea (Channell & Mareschal 1989, Spadini *et al*. 1995) as well as the Valencia Trough (Vegas 1992, Torres *et al*. 1993) went through the main phase of opening (Fig. 34). Rifting in the Pantelleria Trough, between Africa and Sicily occurred in the Pliocene and Quaternary (Golonka 2000, 2002, 2004). A back-arc basin formed in the Aegean area behind the subduction zone. The main folding and thrusting phase occurred in northwestern Africa, along with the formation of nappes (Burollet 1991). Carpathian thrusting progressed east and southeastwards, with a strong element of translation (Golonka *et al*. 2000, 2005b). The thrusting was completed during the Pliocene-Quaternary in the Vrancea Mountains in Romania. The eastward movement of the orogen was related to the movement of Corsica and Sardinia and the subsequent opening of the Ligurian and Tyrrhenian Sea. The extrusion caused by collision of the Apulia plate with Europe could have played a role in the eastward movement of the orogen (Decker & Peresson 1996, Golonka *et al*. 2000, 2005b, 2006).

Paratethys continued its existence in central, eastern Europe and central Asia with mainly marginal marine environment (Fig. 34) and fine clastic and carbonate deposition (Fig. 29) (Dercourt *et al*. 1993, Rögl 1999, Golonka *et al*. 2000, 2006, Il'ina 2000, Nevesskaya *et al*. 2003, Oszczypko 2006, Popov *et al*. 2006, Piller *et al*. 2007). Brown coals were abundant in the Pannonian basin. Continental, fluvial and limnic environments continued in the central European lowlands with fine-grain clastic deposition (Ziegler 1982, 1988, 1990, Piwocki 1998, Reichenbacher 2000, Szynkiewicz 2000, Eissmann 2002, Ford & Golonka 2003, Rasser *et al*. 2008). Rifting continued in Western Europe (Ziegler 1988, 1990, 1992, Wilson & Downes 1991, Bois 1993, Wilson 1994, Schäfer *et al*. 2005).

During Neogene times, major uplift of the Norwegian mainland occurred (Golonka 2000, 2002, Japsen & Chalmers 2000, Rohrman *et al*. 2002, Golonka *et al*. 2003b, Rasmussen *et al*. 2008). A thick, basinward-thickening wedge of Plio-Pleistocene fine-grained clastic sedimentation resulted from this uplift (Ziegler 1988, 1990, 1992, Golonka 2000, 2002, Eidvin *et al*. 2000, Laberg *et al*. 2005, Rasmussen *et al*. 2008). A thick wedge of the fine-grained sediments can also be expected in the eastern part of the East Greenland shelf (Larsen 1990, Chalmers 2000, Johnson & Gallagher 2000). According to Solheim *et al*. (1998) glacial deposition has taken place on the East Greenland margin at least since 7 Ma, but apparently only since 2.5 Ma on the Svalbard-Barents Sea margin. Sedimentation of fine grained clastics and sands continued in North Sea main basin, while coarse-grained braided fluvial deposits still were deposited like in previous time slices south of Scandinavia (Berstad & Dypvik 1982, Ziegler 1988, 1990, 1992, Jordt *et al*. 1995, 2000, Eidvin & Rundberg 2007, Japsen *et al*. 2008, Rasmussen *et al*. 2008, Marcussen *et al*. 2009).

Large volumes of sediments were deposited in the south-eastern part of the Barents Sea during Pliocene-Pleistocene times (Fig. 34). Marine seaways and area of shallow marine environment with fine-grained clastic deposition in the Kara Sea and southern Barents Sea basin enlarged (Meyerhoff 1983, Green *et al*. 1984, Ronov *et al*. 1989, Bogatski *et al*. 1996, Volkova & Kul'kova 1996, Kulkova & Volkova 1997, Tull 1997, Bogdanov *et al*. 1998, Musatov & Pogrebitskij 2000, Akhmet'ev *et al*. 2001, Golonka *et al*. 2003b, Kuz'mina & Volkova 2008). Transgression advanced mainly across the Laptev, East Siberian and the Chukchi seas (Fotina 1994). Over 10 km of Cenozoic deposits was recorded on a seismic secton in the Laptev Sea rift, which is an extension of Eurasian Basin (Drachev *et al*. 1998, Franke *et al*. 2000).

Spreading between Africa, South America and Antarctica (Fig. 35) continued from the previous time slices (Lawver *et al*. 1985, Nürnberg & Müller 1991, Lawver & Gahagan 1993, Golonka 2000, 2002, Golonka & Ford 2003, Macdonald *et al*. 2003). Spreading in the South Atlantic occurred more rapidly then in the North Atlantic (Müller *et al*. 1997, Nürnberg & Müller 1991). Scotia arc moved further eastward, reaching east-west spreading ridge in this area. Right lateral strike-slip fault between Scotia Sea and Antarctic peninsula was active (Lawver *et al*. 1992, Maldonado *et al*. 1998, 2005, Golonka 2000, 2002, Barker 2001, Eagles *et al*. 2005). Mixed clastics were deposited on the South American and Antarctic margins while carbonates were abundant around the South African margin (Dingle & Hendry 1984, 1989, Clemson *et al*. 1997, Billups, *et al*. 2002, Richardson & Underhill 2002, Ford & Golonka 2003, Kiessling *et al*. 2003, Maldonado *et al*. 2005, Wigley & Compton 2006, Parras *et al*. 2008). Deltas of Orange and Limpopo rivers continue supply of coarse clastics (Goudie 2005, Bluck *et al*. 2007). Marine basins with clastic sedimentation rich in glauconite existed in Argentina (Pindell & Tabbutt 1995, Ford & Golonka 2003). Volcanics and volcanoclastic were abundant in the Andean foreland (Giacosa & Heredia 1987, Bissig *et al*. 2002, Ford & Golonka 2003, Gorring *et al*. 2003, Lagabrielle *et al*. 2004, Martina *et al*. 2006). Deposition of diatomaceous ooze and fine-grained clastics with silica, rich in radiolarians continued on Falkland Plateau (Fig. 35) (Muza & Wise 1983, Weaver 1983, Howe *et al*. 1997, Richardson & Underhill 2002, Ford & Golonka 2003).

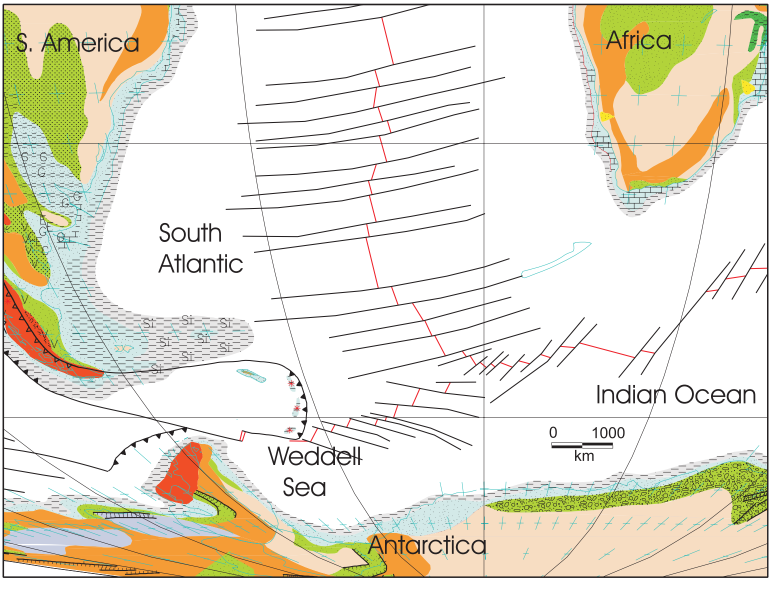

Fig. 35. Plate tectonic, paleoenvironment and lithofacies map of South Atlantic, southwestern Indian Ocean and adjacent areas during Late Miocene-Pliocene times

Fig. 35. Mapa tektoniki płyt, paleośrodowiska i litofacji południowego Atlantyku, południowo--zachodniego Oceanu Indyjskiego oraz obszarów sąsiednich w pliocenie

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Streszczenie

Artykuł przedstawia szczegółowe mapy paleogeograficzne dla siedmiu przedziałów czasowych w obrębie kenozoiku. Trzydzieści pięć map zawiera informacje dotyczące tektoniki płyt, paleośrodowiska i paleolitofacji w czasie paleocenu, eocenu, oligocenu, miocenu i pliocenu. Mapy były skonstruowane przy użyciu programów PLATES i PALEOMAP. Wulkanity znaczące działalność plam gorąca pomagają przy określaniu paleodługości geograficznej. Dane paleomagnetyczne posłużyły do oznaczenia paleoszerokości geograficznej. Informacje zawarte w szeregu globalnych i regionalnych prac zostały wyselekcjonowane i naniesione na mapy. Mapy paleogeograficzne ilustrują geodynamiczną ewolucję Ziemi od późnej kredy po neogen, rozrost (spreding) i tworzenie się oceanów, zamykanie się oceanów, kolizje, łączenie się kontynentów i tworzenie się nowych superkontynentów. Przestrzenną rekonstrukcję architektury basenów w okresie ich powstawania, ekspansji, zamykania i inwersji oraz analizę dynamiki grzbietów śródbasenowych uzyskano, wykonując modelowanie palinspastyczne, przy uwzględnieniu badań paleomagnetycznych oraz analizy stratygraficzno-facjalnej basenów i rozdzielających je grzbietów. W obrębie basenów wydzielono poszczególne strefy paleośrodowiskowe i paleolitofacjalne.

W okresie późnej kredy-wczesnego paleocenu następuje wyraźne rozszerzenie Atlantyku równikowego (Fig. 1, 2). Podniesione kontynenty wokół oceanu dostarczały znacznych ilości materiału klastycznego. Istniały rozległe platformy węglanowe między innymi na obszarze wysp Bahama, Florydy i Jukatanu. Trwały procesy tektoniczne w obszarze karaibskim i w Andach w Ameryce Południowej. Płyta indyjska była ulokowana na półkuli południowej i zbliżała się w kierunku Eurazji (Fig. 3). Afryka i Arabia zbliżała się również do Eurazji, na skutek rozszerzania się Atlantyku południowego (Fig. 3, 5). Kolizja pomiędzy jednostkami austroalpejskimi i teranem briansońskim w Alpach rozpoczęła się w paleocenie (Fig. 4). W późnej kredzie-najwcześniejszym paleocenie zamknął się basen pienińskiego pasa skałkowego i miała miejsce kolizja Karpat Wewnętrznych z grzbietem czorsztyńskim. Złożony system fałdów i nasunięć rozwinął się w pienińskim pasie skałkowym. Wynikiem podnoszenia się obszarów krawędziowych była ogromna ilość materiału klastycznego dostarczanego do basenów fliszowych. Trwał rozrost północnego Oceanu Atlantyckiego. Morza Labradorskiego i jego przedłużenia, basenu Makarowa w Arktyce (Fig. 4).

W okresie paleocenu-środkowego eocenu (Fig. 6–10) trwał dryft Ameryki Północnej i Południowej. Z dryftem tym związane były procesy tektoniczne w Andach, Górach Skalistych i na obrzeżach Morza Karaibskiego w Ameryce Południowej. Procesom tym towarzyszył rozwój basenów sedymentacyjnych. Trwała sedymentacja węglanowa na platformach wysp Bahama, Florydy i Jukatanu (Fig. 7).

Mapy Azji południowej (Fig. 6, 8) przedstawiają tradycyjny punkt widzenia dotyczący kolizji Indii z Eurazją. Ocean Indyjski utworzył się w wyniku ruchu Indii w kierunku północnym. Kolizja nastąpiła w eocenie, rozpoczynając proces tworzenia się Himalajów. Orogeneza himalajska miała istotny wpływ na rozwój tektoniczny Azji południowo-wschodniej, przyczyniając się w trzeciorzędzie do powstania licznych basenów ekstensyjnych. W paleocenie-eocenie trwało zamykanie Neotetydy przez orogenezy alpejską i himalajską. Płyty Adria (Apulia), Alp Wschodnich (austroalpejska) i blok Karpat Wewnętrznych nieustannie przesuwały się ku północy (Fig. 9). Ich kolizja z płytą europejską zaczęła się w Alpach około 47 milionów lat temu. Blok Alkapa powstał przez złączenie Alp Wschodnich, Karpat Wewnętrznych oraz mniejszych płyt, takich jak Bükk, bloki transdunajski czy bukowińsko-getycki. Zamknięcie basenu pienińskiego pasa skałkowego w Karpatach zostało zakończone i rejon ten został włączony do pryzmy akrecyjnej basenu magurskiego. W późnym eocenie nastąpiło również fałdowanie strefy renodunajskiej. Północny Atlantyk i Morze Grenlandzkie otworzyły się w paleogenie (Fig. 9). W tym czasie trwał też rozrost południowego Oceanu Atlantyckiego (Fig. 10).

W okresie późnego eocenu (Fig. 11–15) trwała kontynuacja procesów tektonicznych i sedymentacyjnych wokół Atlantyku centralnego, Morza Karaibskiego, Zatoki Meksykańskiej, w Andach i w Górach Skalistych (Fig. 12). Kolizja Indii z Eurazją spowodowała tektonikę ucieczki Azji południowo-wschodniej, między innymi powstanie uskoku Rzeki Czerwonej (Fig. 13). Otwarcie północnego Atlantyku było związane z działalnością plamki gorąca i pióropusza płaszcza Islandii (Fig. 14).

W oligocenie (Fig. 16–20) otwarciu się basenów oceanicznych wokół Antarktydy, towarzyszyło powstanie lodowców na tym kontynencie (Fig. 20). W tym czasie trwały też kolizje (Fig. 18, 19) pomiędzy Indiami, Afryką i Eurazją. Doszło do kolizji Apulii jak również teranów alpejsko-karpackich z płytą europejską. Morze Paratetydy powstało w Europie i Azji centralnej, przed posuwającymi się na północ pasmami orogenicznymi. Geodynamiczna ewolucja basenów w paśmie alpejskim doprowadziła do przejścia typu sedymentacji od fliszu do molasy. Procesy tektoniczne i sedymentacyjne wokół Atlantyku centralnego, Morza Karaibskiego, Zatoki Meksykańskiej, w Andach i w Górach Skalistych (Fig. 17) były kontynuacją tych z poprzedniego okresu. Trwała sedymentacja węglanowa na platformach wysp Bahama, Florydy i Jukatanu (Fig. 17).

We wczesnym miocenie (Fig. 21–25) następowało dalsze rozszerzanie się oceanów Atlantyckiego i Arktycznego, a także dryft obu Ameryk. Procesy tektoniczne w rejonie Morza Karaibskiego i północnej części Ameryki Południowej doprowadziły do powstania przesmyku panamskiego, ciągle zalanego przez morze (Fig. 22). Istniało połączenie morskie pomiędzy Atlantykiem i Pacyfikiem. Szat-burdygał był okresem głównej fazy orogenezy alpejskiej, formowania się gór w obszarze alpejsko-karpackim, śródziemnomorskim, himalajskim i Azji centralnej (Fig. 23, 24). Kolizja Australii i Filipin z Eurazją w neogenie zapoczątkowała tektonikę kompresyjną w Azji południowo-wschodniej. W zachodniej części Tetyda została zastąpiona przez współczesne Morze Śródziemne. Arabia oddzieliła się od Afryki, powstało Morze Czerwone i ryft wschodnioafrykański (Fig. 23, 24). Ruchy przesuwcze w rejonie Grenlandii i Svalbardu doprowadziły do szerokiego połączenia północnego Atlantyku z basenem eurazjatyckim w Arktyce. Zaawansowany spreding trwał na obszarze pomiędzy Ameryką Południową, Afryką i Antarktydą (Fig. 25).

W środkowym miocenie (Fig. 26–30) nastąpiło przyśpieszenie ruchu Ameryki Południowej. Związane z tym było odnowienie ruchów orogenicznych miocenie Andach i powstanie strefy subdukcji na południowym obrzeżeniu Morza Karaibskiego (Fig. 27, 30).

Orogeneza himalajska w Azji związana jest dalszym procesem ekstruzji Azji południowo-wschodniej, rozwojem uskoków przesuwczych i związanych z nimi basenów, których przykładem jest basen Rzeki Czerwonej w Wietnamie (Fig. 28). W okresie tym powstało Morze Południowochińskie. Kolizja Arabii z płytami irańskimi doprowadziła do powstania orogenu gór Zagros. Kolizje Indii i Arabii miały swój wpływ na geodynamiczną ewolucję Azji centralnej, a także Kaukazu, Morza Czarnego i Morza Kaspijskiego. Na obszarze alpejsko-karpackim nastąpiła główna faza kolizji Alkapy i Adrii z płytą europejską. Ukośna kolizja pomiędzy płytą północnoeuropejską i najeżdżającymi na nią płytami Karpat Zachodnich prowadziła do rozwoju zewnętrznej pryzmy akrecyjnej, uformowania się szeregu płaszczowin fliszowych i utworzenia zapadliska przedgórskiego. Płaszczowiny były odkłute od swojego pierwotnego podłoża i nasunięte na paleozoiczno-mezozoiczne osady platformy północnoeuropejskiej pokryte częściowo przez utwory trzeciorzędowe. Utwory te na Niżu Europejskim zawierają facje środowiska jeziornego z pokładami węgla brunatnego (Fig. 29).

Zaawansowany spreding trwał na obszarze pomiędzy Ameryką Południową, Afryką i Antarktydą (Fig. 30). Rozwijało się zlodowacenie na półkuli południowej.

W późnym miocenie-pliocenie (Fig. 31–35) nastąpił dalszy rozwój orogenezy andyjskiej i reorganizacja tektoniczna rejonu Morza Karaibskiego. W późnym miocenie ciągle istniało połączenie morskie pomiędzy Atlantykiem i Pacyfikiem (Fig. 32). Zostało ono zlikwidowane w pliocenie, kiedy przesmyk środkowoamerykański (panamski) stał się obszarem lądowym. Zamknięcie przesmyku spowodowało zmianę cyrkulacji oceanicznej i w konsekwencji zmianę klimatu na chłodniejszy oraz powstanie zlodowacenia na półkuli północnej. Klimat Ziemi w kenozoiku odzwierciedla etapy tektoniczne rozpadu i łączenia się kontynentów. Klimat zmieniał się od ciepłego do współczesnego okresu zimnego. Przez cały paleogen i neogen aż po pliocen istniały rozległe platformy węglanowe na obszarze wokółatlantyckim, między innymi na obszarze wysp Bahama, Florydy i Jukatanu (Fig. 32). Zatoka Meksykańska natomiast charakteryzowała się sedymentacją klastyczną płytkomorską na obrzeżach i głębokomorską w centralnej części basenu.

Rejon obrzeży Oceanu Indyjskiego charakteryzuje powstanie licznych delt (Fig. 33). Delty Indusu i Gangesu i ich stożki sięgające daleko w głąb oceanu były zasilane materiałem z wypiętrzanych i erodowanych Himalajów. Himalaje dostarczały też materiał klastyczny do położonego w północnych Indiach rowu przedgórskiego. Kolizja Indii z Eurazją doprowadziła do powstania uskoków przesuwczych sięgających w głąb Azji centralnej (Fig. 34). Na obszarze pomiędzy Afryką a Europą w wyniku ewolucji morza Alboran nastąpiło zamknięcie Cieśniny Gibraltarskiej. Odcięcie Morza Śródziemnego od Atlantyku spowodowało kryzys salinarny. Seria ewaporatów o miąższości 3000 m osadziła się na dnie morza. Ruchy górotwórcze miały miejsce w północnej Afryce. W północnej Europie miało miejsce wypiętrzanie Norwegii oraz związana z tym erozja i dostarczanie materiału klastycznego do basenów północnoatlantyckich (Fig. 34). Zaawansowany spreding trwał w dalszym ciągu na obszarze pomiędzy Ameryką Południową, Afryką i Antarktydą (Fig. 35). Delty rzeki Orange i Limpopo dostarczały materiał klastyczny do basenów wokół południowej Afryki.