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The evolution of late Ediacaran riverine-estuarine system in the Lublin–Podlasie slope of the East European Craton, southeastern Poland

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CONTENTS

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Estuaries are unlike other coastal systems, however, because they are geologically ephemeral: if the rate of sediment supply is sufficient, then estuaries become filled and cease to exist when the rate of sea-level rise slows

Robert W. Dalrymple, Brian A. Zaitlin, Ron Boyd (1992)

Jolanta PACZEŚNA – **The evolution of late Ediacaran riverine-estuarine system in the Lublin–Podlasie slope of the East European Craton, southeastern Poland.** *Polish Geological Institute Special Papers,* 27: 1–96

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Abstract. A model of evolution and destruction of the late Ediacaran riverine-estuarine system that developed in the Podlasie Depression and Lublin–Podlasie slope of the East European Craton is presented based on the identification and definition of facies associations, depositional systems and the framework of high-resolution sequence stratigraphy. Two groups of depositional systems have been identified – alluvial, estuarine, and one open coast system. The alluvial system was initially represented in the northeastern and western synrift depocentres by alluvial fans. Distal parts of the fans were areas of fluvial deposition. Large, sand-bed braided rivers flowed transverse to the sedimentary basin axis. During final stages of the alluvial basin evolution, the levelling of the rift topography and the increase in subsidence rate in the south-east of the basin resulted in the development of anastomosed system rivers. The rivers flowed along the basin axis from north to south. A change in the braided-river flow type from ephemeral during the early stage of the alluvial basin evolution to perennial in the later stages, development of anastomosed system river floodplains and the change in the colour of accumulated fluvial deposits indicate a climate change from arid and desert to more humid and moderate conditions. The late Ediacaran siliciclastic succession of the Lublin basin is a record of the transgressive stage of estuary development. It is manifested by five successive parasequences composing the transgressive systems tract. During the earliest evolutionary stages, the Lublin estuary was a mixed wave-and tide-dominated. In its peak development, as the influence of tides significantly increased, it turned into a macrotidal, hypersynchronous estuary of funnel-shaped geometry. Regression of the Lublin estuary, resulting in its ultimate decline, started along with the highstand development at the Ediacaran/Cambrian transition. The estuary became transformed into a mixed-energy wave-and tide-dominated estuarine system and subsequently into a wave-dominated open coast.

Key words: sedimentology, sequence stratigraphy, alluvial plain, estuary, late Ediacaran, early Lower Cambrian, East European Craton, southeastern Poland.

Abstrakt. Na podstawie zdefiniowania asocjacji facjalnych i systemów depozycyjnych oraz przedstawienia ram wydzieleń wysokorozdzielczej stratygrafii sekwencji sformułowano model rozwoju i destrukcji późnoediakarskiego systemu fluwialno-estuariowego, rozprzestrzenionego w obniżeniu podlaskim i lubelskim skłonie kratonu wschodnioeuropejskiego. Wyróżniono dwie grupy systemów depozycyjnych – aluwialne i estuariowe oraz system otwartego wybrzeża. System aluwialny był początkowo reprezentowany w północno-wschodnich i zachodnich depocentrach synryftowych przez stożki aluwialne. Dystalne części stożków były obszarami depozycji fluwialnej. Duże, piaskodenne rzeki roztokowe spływały poprzecznie do osi basenu sedymentacyjnego. W końcowych stadiach ewolucji basenu aluwialnego wyrównanie topografii ryftowej i wzrost tempa subsydencji w jego południowo-wschodniej części spowodowały rozwój rzek systemu anastomozującego. Spływały one wzdłuż osi basenu z północy na południe. Zmiana rodzaju przepływu rzek roztokowych z okresowego we wczesnych etapach rozwoju basenu aluwialnego na ciągły w późniejszych stadiach, rozwój równi zalewowych rzek systemu anastomozującego i zmiana koloru osadów akumulowanych przez rzeki świadczą o zmianie klimatu suchego, pustynnego na bardziej wilgotny, umiarkowany. Późnoediakarska sukcesja silikoklastyczna basenu lubelskiego jest zapisem transgresywnego etapu ewolucji estuarium. Jej przebieg odzwierciedla pięć kolejnych parasekwencji budujących transgresywny ciąg systemowy. W najwcześniejszych etapach rozwoju estuarium lubelskie miało charakter mieszany, falowo-pływowy. W fazie maksymalnego rozwoju, w miarę znaczącego wzrostu oddziaływania pływów, było to makropływowe, hypersynchroniczne estuarium o kominowej geometrii. Na przełomie ediakaru i kambru wraz z rozwojem ciągu systemowego wysokiego stanu względnego poziomu morza rozpoczął się regres i stopniowa likwidacja estuarium lubelskiego, które przekształciło się w estuarium o mieszanej energii falowo-pływowej i następnie w otwarte wybrzeże z udziałem falowania.

Słowa kluczowe: sedymentologia, stratygrafia sekwencji, równia aluwialna, estuarium, późny ediakar, wczesny dolny kambr, kraton wschodnioeuropejski, południowo-wschodnia Polska.

INTRODUCTION

The Ediacaran Period was marked in the Earth's history by a number of important global-scale geological, physical, geochemical and biological events.

Geotectonic processes that occurred during the final stage of breakup of the Rodinia supercontinent resulted in a dispersion of continental plates (e.g. Cawood *et al*., 2007; Elming *et al*., 2007; Li *et al*., 2008). The dispersion process started in the Cryogenian and reached its final stage during the Ediacaran, manifesting itself by the rifting phase followed by a drift of continental plates. At the Ediacaran/Cambrian boundary, passive margins developed along previously tectonically active margins, including western Baltica.

Global glaciations, followed by a deglaciation period, were the prominent geological events that subsequently affected the Ediacaran atmosphere, hydrosphere and biosphere, and found their record in sedimentary processes (Rieu *et al*., 2007a). The two Neoproterozoic younger glaciations occurred after the main stages of Rodinia breakup in the Cryogenian: Marinoan about 635 Ma and Gaskiers at ca. 580 Ma. The last Neoproterozoic glaciation may have occurred during the early Ediacaran (Fairchild, Kennedy, 2007). Its effect was manifested by an increase in oxygen content in the atmosphere. Ice-sheet melting resulted in a rise of sea level (Rieu *et al*., 2007b). Climatic variability during Ediacaran times caused some perturbations in geochemical cycles recorded by fluctuations in isotopic contents of organic carbon and strontium (e.g. Hoffan, Schrag, 2002; Halverson *et al*., 2005; LeGuerroue *et al*., 2006).

As the sea level rose, most of the Ediacaran continents, including Baltica, became inundated by marine transgressions. Increased oxygenation of the atmosphere and hydrosphere as well as the development of new marine ecosystems resulted in unprecedented changes in the Ediacaran biosphere, unknown before in the Earth's history. The changes stimulated the appearance of widespread Ediacaran-type faunas representing soft-bodied Metazoa and new species of microorganisms represented by unicellular eukaryotes and prokaryotes. Among these organisms there were ancestors of some Palaeozoic phyla (e.g. McCall, 2006) and some species of microplankton, which have been recorded in recent years (Moczydłowska, 2008a). The organic spectrum was clearly characterised at that time by an increase in benthic species boring morphologically simple feeding-dwelling burrows in the sediment (e.g. Fedonkin, 1977; Pacześna, 1986; Crimes, 1994; Crimes, Fedonkin, 1996).

The above mentioned geological, physical, geochemical and biological factors left global (Fig. 1) and regional records in Ediacaran siliciclastic deposits of the East European Craton. Their effects are observed in the evolution of the late Ediacaran sedimentary basin in the Lublin slope of the East European Craton and Podlasie Depression (e.g. Pacześna, 2008; Moczydłowska, 2008b).

The principal objective of this report is to document and present a depositional model and constraints on the development of a hypothetical estuary that evolved during the late Ediacaran in the Lublin slope of the East European Craton. The following undertakings were made to achieve the target:

– defining and describing the late Ediacaran sedimentary environments, post-volcanic siliciclastic deposits;

– applying a high-resolution sequence stratigraphic method to study the post-volcanic siliciclastic basin fill;

– mapping the distribution of individual sedimentary environments within the basin in time and space;

– comparing the Ediacaran sedimentary model of a hypothetical estuary with the Phanerozoic and modern equivalents, assuming that the quality, quantity and range of the effects of geological, physical, geochemical and biological factors which affected sedimentary processes during the Neoproterozoic times were different from those operating today.

Although the Ediacaran System is the final link in the Precambrian evolution of the Earth, its unique geological characteristics impede simple comparisons of the Ediacaran sedimentation of siliciclastic deposits with the Phanerozoic and recent models of analogous depositional systems (Eriksson *et al*., 1998). It was due mainly to different factors that operated on a global scale – much higher sedimentation rate and the Earth-Moon dynamic phases recorded in tidal settings, which were different from the recent ones, indicating different lengths of calendar days, months and years (Williams, 1998). The other factors included the following: different physical and chemical composition of the atmosphere and hydrosphere, lack of plant cover and increased chemical and mechanical weathering due to higher CO_2 content in the air (Fike *et al*., 2006).

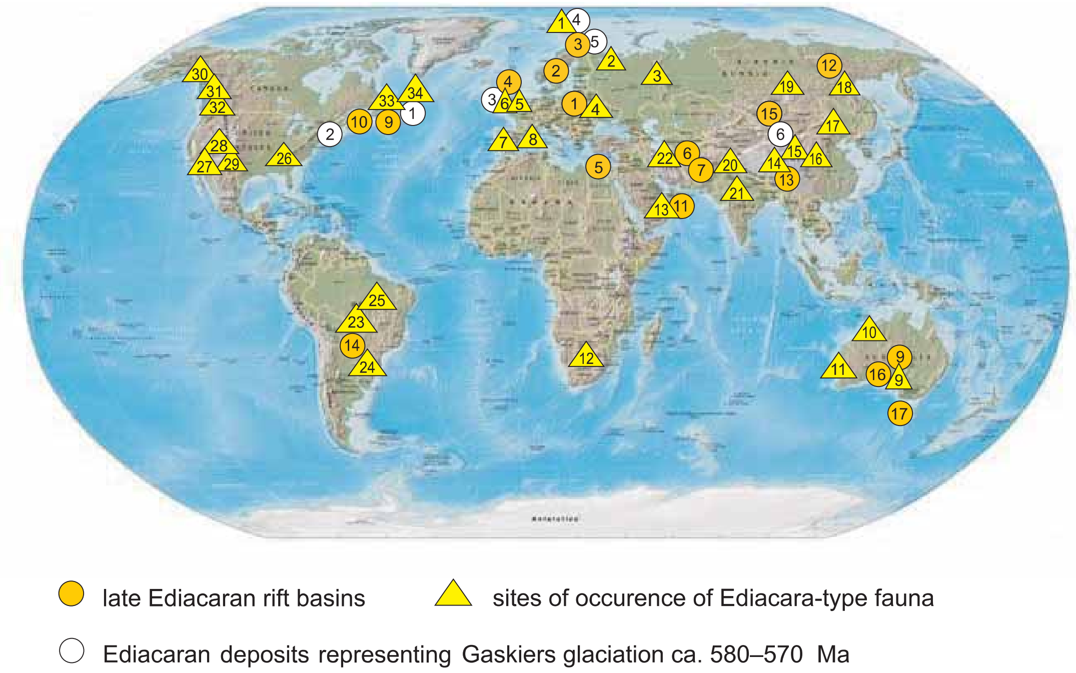

Fig. 1. Occurrences of the most important Ediacaran geological and palaeobiological events

Late Ediacaran rift basins: 1 – Lublin–Podlasie basin – Poland, Volhyn and Podolia – Ukraine, 2 – Muhos basin, Sweden, 3 – Vättern graben, Norway, 4 – northern Scotland, 5 – Sinai Peninsula, 6 – central Iran, 7 – Salt-Range Province, Pakistan, 8 – central superbasin (Georgina basin, Amadeus basin, Officer basin), Australia, 9 – south-eastern New Foundland, Canada, 10 – New Brunswick, Canada, 11 – southern Oman, 12 – Lena, Tunguska province, Russia, 13 – southern China, 14 – north-eastern Argentina, 15 – Zavkhan basin, south-western Mongolia, 16 – Adelaide rift complex, southern Australia, 17 – western Tasmania, Australia

Occurrences of Ediacaran-type soft-bodied Metazoa fauna: 1 – Scotia Group, Princ Karls Foreland, Svalbard, 2 – Winter Coast, White Sea, Russia, 3 – Urals, Russia, 4 – Podolia, Ukraine, 5 – Charnwood Forest, England, 6 – South Wales, England, 7 – Galicia and central Spain, 8 – Sardinia, 9 – southern Australia, Ediacara, Flinders Range, 10 – Northern Territory, Amadeus Basin, Australia, 11 – Kimberleys and Stirling Range, Western Australia, 12 – Namibia, South-West Africa, 13 – Huqf Group, southern Oman, 14 – Wengan, southern China, 15 – Huainan, China, 16 – Yangtze Gorges, China, 17 – Liaodong, eastern China, 18 – Jixi, eastern China, 19 – south-western Mongolia, 20 – Knoll Formation, Lesser Himalaya, India, 21 – Bhander Group, Vindhyan Supergroup, central India, 22 – northern Iran, 23 – north-western Argentina, 24 – Yerbal Formation, Arroyo del Soldado Group, south-eastern Uruguay, 25 – Paraguay River Basin, Brazil, 26 – northern Carolina, United States, 27 – California: Salt Spring – southern Death Valley, Kelso Mountains – Mojave Desert, White Mountains – eastern California, United States, 28 – southern Nevada, 29 – Clemente Formation, Mexico, 30 – North West Territories, Canada, 31 – Mackenzie Mountains, Yukon, Canada, 32 – Rocky Mountains, British Columbia, Canada, 33 – Istaken Point, Avalon Peninsula, New Foundland, Canada 34 – Spaniard's Bay, Avalon Peninsula, New Foundland, Canada

Ediacaran deposits of the Gaskiers glaciation ca. 580 Ma: 1 – New Foundland, Canada, 2 – Massachusetts, United States, 3 – Scotland, England, 4 – north--eastern Svalbard, 5 – northern Norway, 6 – Hankalchough, China

The late Ediacaran deposits of the Lublin slope of the East European Craton and Podlasie Depression have been broadly studied for over 40 years, as they take a key position within the complicated geotectonic evolution of southeastern Poland during the dynamic processes of sedimentary basin filling. Complete or fragmentary Ediacaran successions have been encountered in 34 wells located within two structural units of the Polish sector of the East European Craton: Podlasie Depression and the Lublin slope of the East European Craton (Fig. 2). The Ediacaran sections in most of the wells, in particular those drilled by the Polish Geological Institute, were fully cored commonly with 100% core yield. Due to the large amount of drilling data on the Ediacaran deposits, the Ediacaran succession of southeastern Poland is an excellent geological object for research projects, unique on a global scale.

The late Ediacaran siliciclastics, presented in this paper, overlie a volcanigenic complex. They were dealt with in a number of regional and stratigraphical reports (Znosko, 1965; Areń, 1974, 1978a, b; 1982; Areń, Lendzion, 1978; Areń *et al*., 1979; Moczydłowska, 1988, 1991; Pacześna, 1989; Vidal, Moczydłowska, 1995; Pacześna *et al*., 2005), ichnological studies (Fedonkin, 1977; Pacześna 1985, 1986, 1996), facies investigations (Jaworowski, 1978; Pacześna, 2006, 2007, 2008), tectonic reports (Poprawa, Pacześna, 2002); sequence stratigraphic analyses (Pacześna, Poprawa, 2005a, b), chemostratigraphic isotopic studies (Strauss *et al*., 1997), geochemical investigations (Wichrowska, 1982) and mineralogical and petrographical researches (Juskowiakowa, 1971, 1974, 1978; Wichrowska, 1978, 2007, 2008).

Interpretation of the origin of the late Ediacaran–early Cambrian succession from the Lublin slope of the East European Craton was one of the most interesting and unsolved problem over the last 30 years. The first sedimentological studies were performed mainly to analyse the sedimentary continuity within the Ediacaran/Cambrian transitional deposits to fix the boundary between these systems. The studies

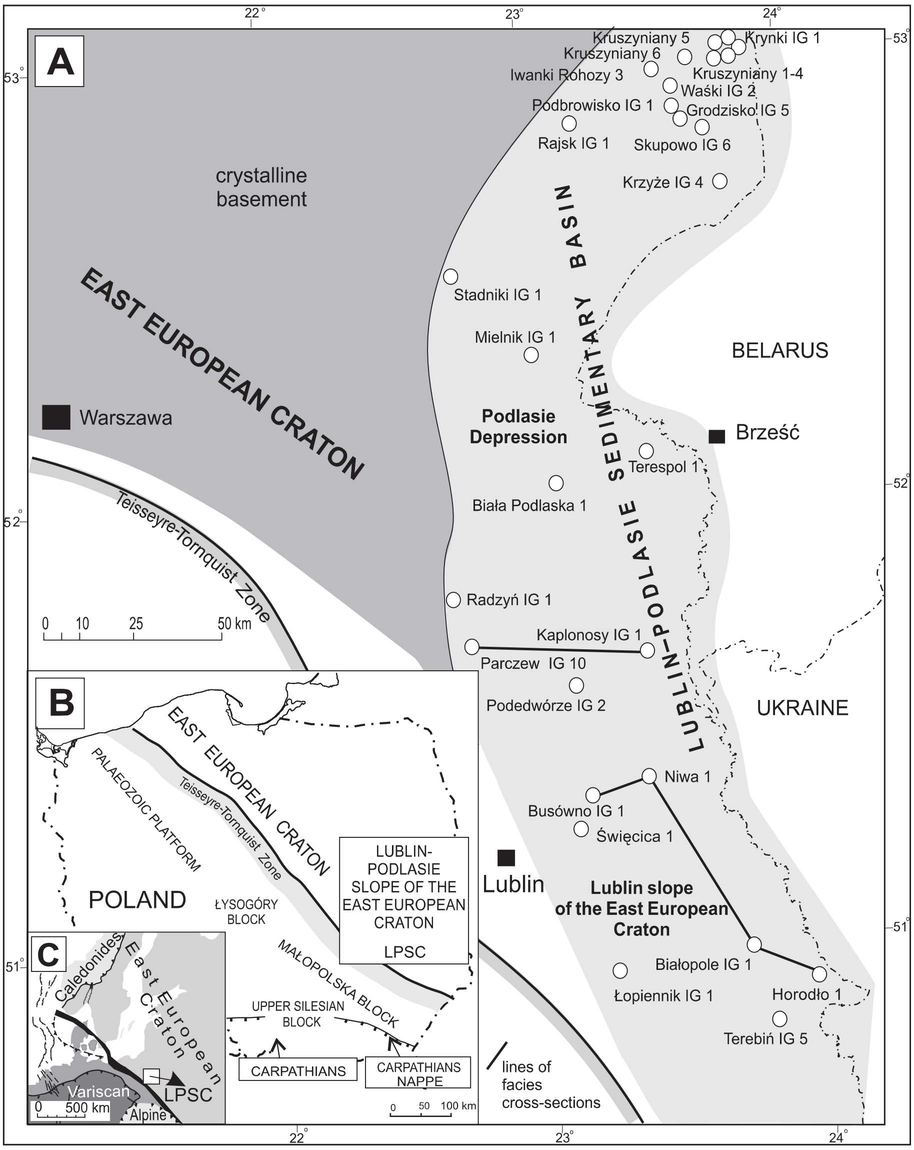

Fig. 2. A – Location of investigated boreholes, B – Location of the Lublin–Podlasie slope of the East European Craton against a sketch-map of tectonic units of Poland, C – Lublin–Podlasie slope of the East European Craton and the main tectonic units of Europe

provided data that enabled general identification of the sedimentary environment representing a shallow epicontinental basin possibly in the tidal zone (Jaworowski, 1978). The Ediacaran/Cambrian transitional deposits from the Lublin slope of the East European Craton were thoroughly investigated for ichnological evidence to determine ichnocoenoses, their ecological structure and relationships with environmental factors (Pacześna, 1996).

GEOLOGICAL SETTING

The development of late Ediacaran sedimentary basins in the western and southwestern margin of the East European Craton, including the Lublin–Podlasie basin, occurred during the final stage of supercontinental cycle representing the Rodinia breakup. The Rodinia supercontinent started to form in the late Mesoproterozoic ca. 1.2–1.0 Ga (Cawood *et al*., 2007). During the Cryogenian, at ca. 800–700 Ma, Rodinia started to dismember into smaller continental blocks. The East European Craton formed the core of paleocontinent Baltica after the breakup of Rodinia. In the present paper the term Baltica is used in a wide terrane sense after Bogdanova *et al*. (2008).

Tectogenesis of the Lublin–Podlasie basin is associated with the final stages of rifting which occurred in western Baltica (East European Craton) during late Neoproterozoic times about 600–550 Ma (Kumpulainen, Nystuen, 1985; Vidal, Moczydłowska, 1995; Andréasson *et al*., 1998; Greiling *et al*., 1999; Poprawa *et al*., 1999; Poprawa, Pacześna, 2002; Jaworowski, Sikorska, 2003; Šliaupa *et al*., 2006; Elming *et al*., 2007).

At those times, Baltica, Amazonia and Laurentia probably separated from one another and the process of Rodinia breakup finished (Cawood *et al*., 2007; Pease *et al.,* 2008). As the result, two systems of rift basins developed in western Baltica during the Ediacaran. One of them was represented by NW–SE-trending basins and was located along the western margin of Baltica. The main stage of its evolution is dated at the late Neoproterozoic (Poprawa, Pacześna, 2002; Jaworowski, Sikorska, 2003; Pharaoh *et al.,* 2006; Šliaupa *et al.,* 2006), however it presumably developed along pre-existing Cryogenian zones (Pharaoh *et al*., 2006). The other basin was represented by the Orsha–Volhyn aulacogen (Pożaryski, Kotański, 1979; Poprawa, Pacześna, 2002), also called the Volhyn–Central Russian aulacogen (Elming *et al*., 2007). It trended NE–SW, obliquely to the western edge of Baltica, and was one of the elements of the rift basin system situated in the central part of the East European Craton between its western and eastern flank (Vidal, Moczydłowska, 1995; Pease *et al.,* 2008). The Ediacaran basin of the Orsha–Volhyn aulacogen developed along pre-existing rift structures (Elming *et al*., 2007) dated at 1.6–0.8 Ga (Bogdanova *et al*., 2008). The late Ediacaran Lublin–Podlasie sedimentary basin was situated at the crossing of the two above mentioned rift basins (Poprawa, Pacześna, 2002) (Fig. 3).

The early major stages of Rodinia breakup and the related rifting event took place in Baltica during the Cryogenian and are dated at approximately 800–700 Ma (Kumpulainen, Nystuen, 1985; Pharaoh *et al*., 2006; Pease *et al*., 2008). They preceded the Gaskiers Glaciation dated at about 580 Ma (Bingen *et al*., 2005). Sedimentary basins that developed as a result of those rifting events occur in north-western Baltoscandia and include the Vättern basin of southern Sweden and the Hedmark and Valdres basins of southern Norway (Pease *et al.,* 2008). The Tanafjorden rifting basin of northern Norway developed on pre-existing late Cryogenian structures. Rifting nature of those basins was related to the Rodinia breakup. Their ages are determined by dating of mafic to ultramafic rock complexes, indicating that the rifting process in Baltica had not finished by approximately 620–550 Ma and may have continued until the Ediacaran/Cambrian transition (e.g. Kumpulainen, Nystuen, 1985; Poprawa, Pacześna, 2002; Cawood *et al*., 2007).

The development of passive continental margin at the western edge of Baltica commenced during early Cambrian times. It coincides with initial phases of the Iapetus Ocean formation (e.g. Siedlecka *et al*., 2004; Cawood *et al*., 2007).

The Cryogenian rifting event in the southwestern marginal zone of Baltica (see Poprawa, Pacześna, 2002) is probably represented by the siliciclastic Polesie Formation deposited in tectonic grabens (Mahnatsch *et al*., 1976). It is observed in

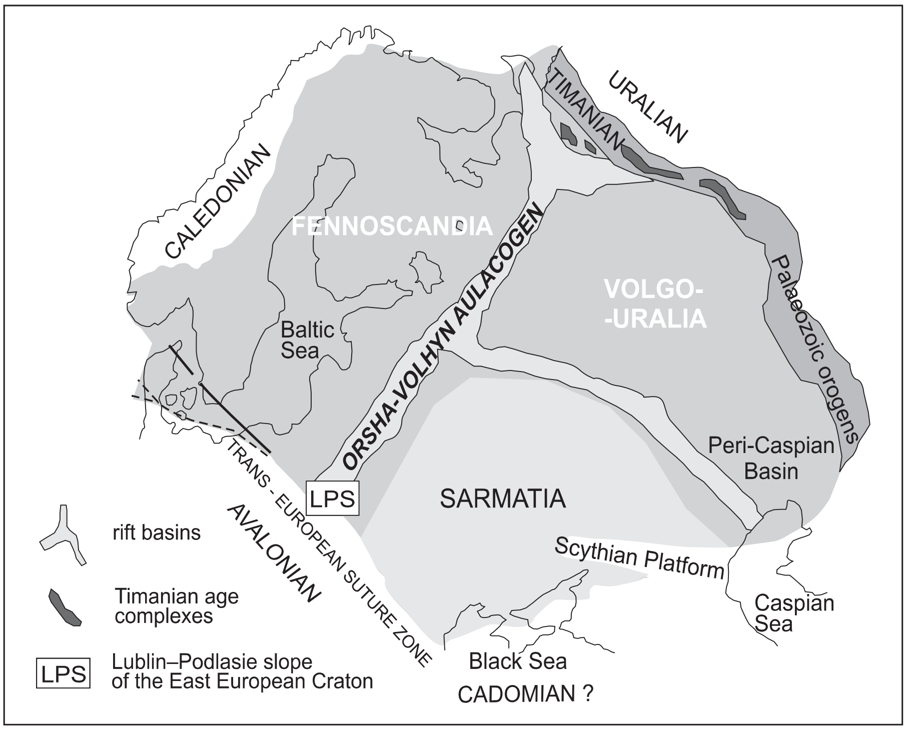

Fig. 3. The late Palaeoproterozoic to Neoproterozoic tectonic complexes of Baltica (East European Craton) (according to Bogdanova *et al*., 2008; Pease *et al*., 2008, simplified)

the Lublin–Podlasie basin under the late Ediacaran sedimentary-volcanigenic complex. The formation represents the lowest 2nd order depositional sequence bounded at top and base by angular unconformities (Pacześna, Poprawa, 2005a, b).

The basin under study is subdivided into the Lublin zone (comprising a structural unit of the Lublin slope of the East European Craton) and the Podlasie zone equivalent to the Podlasie Depression (Fig. 2). Both these zones exhibited different facies development during late Ediacaran to earliest Cambrian times. The facies variability was associated with both the lack of synchronicity between rifting events in these basin zones and unequal development of synrift depocentres (Pacześna, 2006). In both these zones, a non-metamorphosed sedimentary cover rests upon the Palaeoproterozoic and Mesoproterozoic crystalline basement, respectively (e.g. Malinowski *et al*., 2005; Krzemińska *et al*., 2007).

The oldest deposits of the area are represented by the siliciclastic Polesie Formation (Areń, 1982; Poprawa, Pacześna, 2002) (Fig. 4). After a long period of intense erosion, coarse-clastic deposits of the Żuków Formation were deposited above the Polesie Formation (e.g. Juskowiakowa, 1971; Areń, 1982; Poprawa, Pacześna, 2002). The overlying Sławatycze Formation is represented by basalts, tuffs and epiclastic rocks. This sedimentary-volcanigenic complex rests unconformably either upon the Polesie Formation deposits in the central part of the basin or immediately upon the crys-

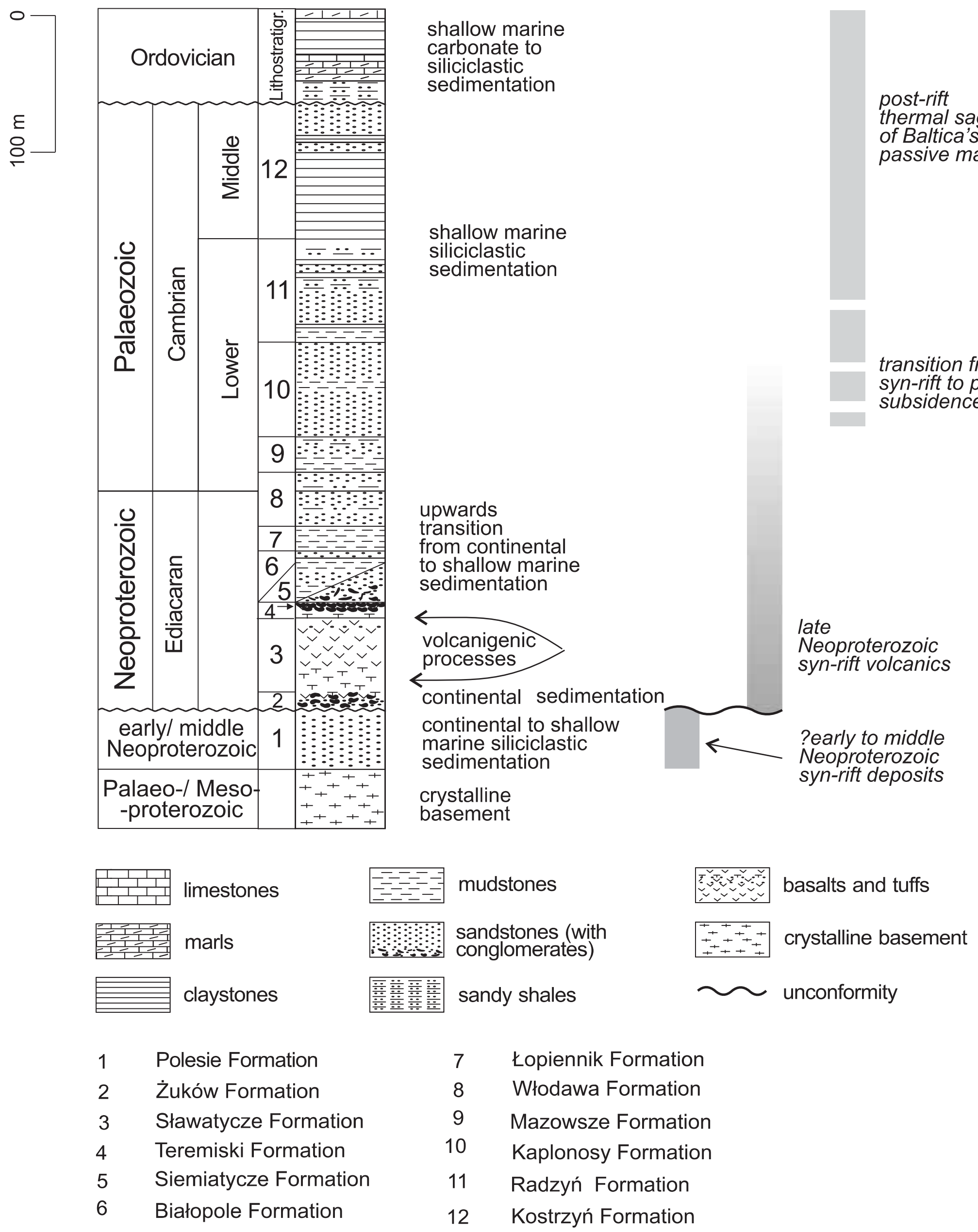

Fig. 4. Main tectonic stages of the Lublin–Podlasie basin evolution and facies development of the basin infill (after Poprawa, Pacześna, 2002, modified)

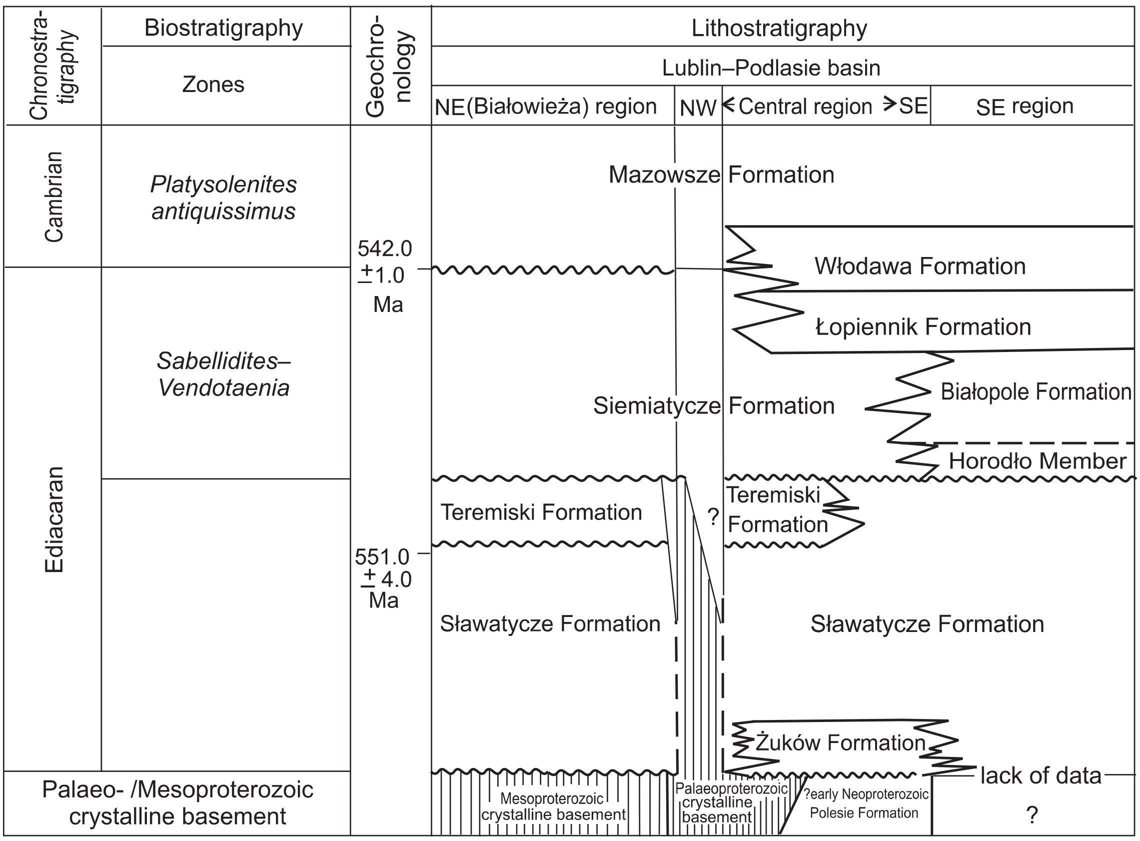

Fig. 5. Chronostratigraphy, biostratigraphy and lithostratigraphy of the late Ediacaran and lowermost Lower Cambrian in the Lublin–Podlasie slope of the East European Craton

Chronostratigraphy after Moczydłowska (1991) and Pacześna (1996); biostratigraphy after Moczydłowska (1991) and Pacześna (1996, 2008); geochronology after International Commission on Stratigraphy (2010); lithostratigraphy after Pacześna (in press)

talline basement in its north-eastern and eastern areas (Fig. 5). Volcanigenic deposits of the Lublin–Podlasie area are a small southwestern part of a large trap basalt province extending outside the territory of Poland towards the east and south-east in western Ukraine, Moldova, Belarus and Russia, covering an area of approximately 200 000 km^2 (Ryka, 1984; Rozanov, Łydka, 1987; Bogdanova *et al.*, 1997; Elming *et al.*, 2007). Rifting nature of the late Ediacaran magmatism was documented by geochemical investigations of basalts, conducted by Białowolska *et al.* (2002) and Krzemińska (2005). Continental basalts infilled the tectonic grabens and half-grabens that developed due to rifting processes during the late Ediacaran final stage of Rodinia breakup (Poprawa, Pacześna, 2002; Pacześna, 2006; Elming *et al.*, 2007). Sedimentary infill of the final stage of synrift phase is represented by continental deposits of the Siemiatycze Formation and shallow marine sediments of the Białopole, Łopiennik and Włodawa Formations (Fig. 5). In the northern part of the Lublin slope of the East European Craton, alluvial deposits of the Siemiatycze Formation directly overlie the Palaeoproterozoic crystalline basement (Malinowski *et al.*, 2005) (Fig. 5).

A depositional record of both the clastic deposits and those accumulated due to rifting magmatic processes manifested itself as the infill of successive synrift depocentres in the late Ediacaran sedimentary basin (Pacześna, 2006).

Except for a number of U–Pb, K/A and Ar/Ar radiometric dating of volcanic deposits, there is no evidence for establishing precise stratigraphic position of the lower part of the Ediacaran succession. The dating provided ages of about 551.0–590.0 Ma (Compston *et al.*, 1995; Velikanov, Korenchuk, 1997; Elming *et al.*, 2007). The first biostratigraphic unit was established in the uppermost Ediacaran. This is the *Vendotaenia–Sabellidites* Zone, established on the base of occurrence of characteristic specimens representing Cyanophyta and small shelly organisms, respectively. The Zone comprises the Białopole and Łopiennik formations and the whole (e.g. Busówno IG 1 section) or only lower part (e.g. Łopiennik IG 1, Białopole IG 1, Terebiń IG 5, Horodło 1 and Niwa 1 sections) of the Włodawa Formation (Moczydłowska, 1991; Pacześna, 2007, 2008) (Fig. 5).

The upper boundary of the Ediacaran System is drawn directly below the first occurrence of the trace fossil *Trichophycus pedum* (Seilacher) within the lower part of the *Platysolenites antiquissimus* Zone representing the first biostratigraphic zone of the Lower Cambrian (Pacześna, 1989, 1996, 2008) (Fig. 5). According to the global standards, this ichnogenus is considered an index taxon of the Lower Cam-

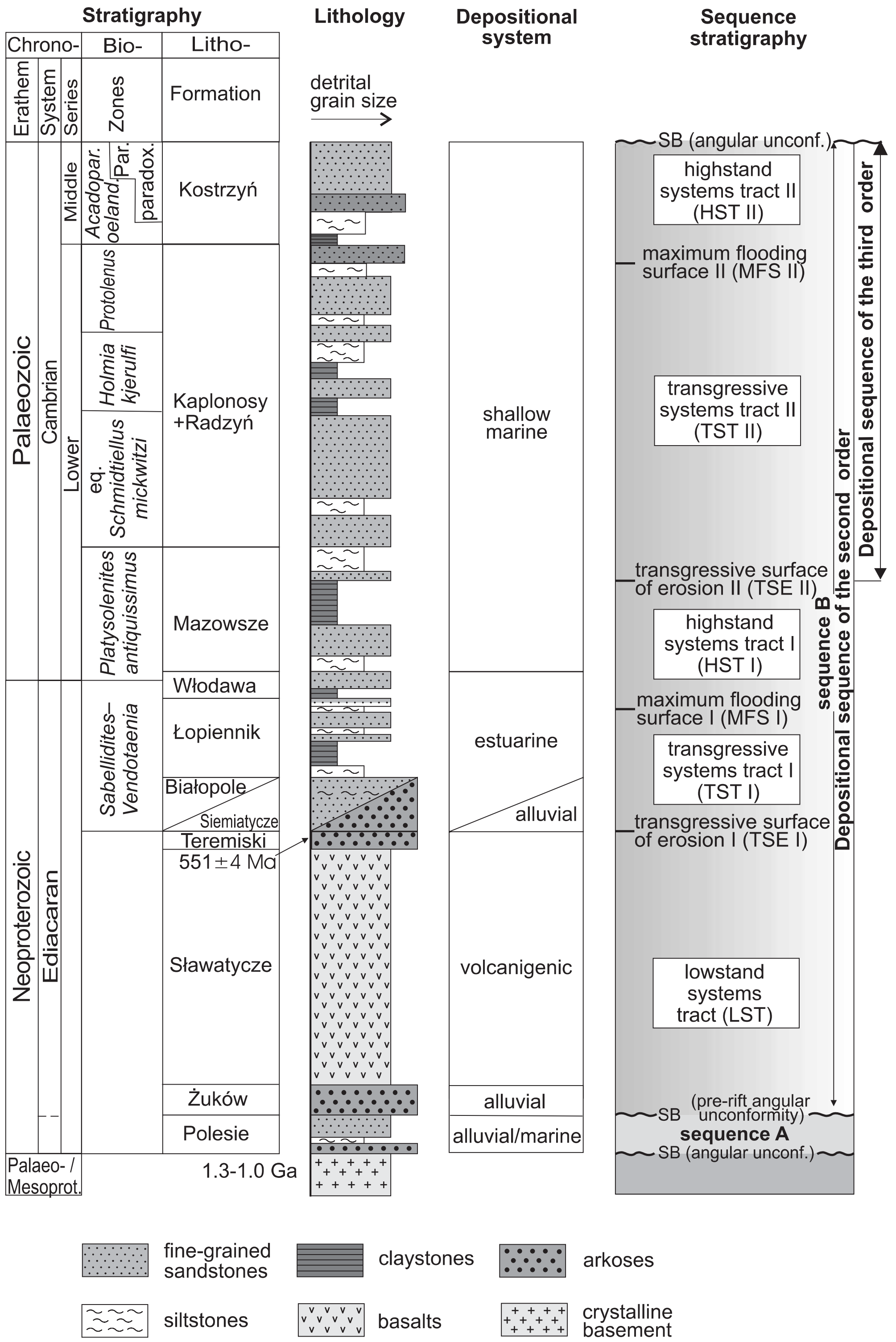

Fig. 6. Ediacaran and Cambrian sequence stratigraphy in the Lublin–Podlasie slope of the East European Craton (after Pacześna, Poprawa 2005a, b)

Biostratigraphy after Lendzion (1983a, b), Moczydłowska (1991) and Pacześna (1996, 2008); Ediacaran lithostratigraphy after Pacześna (in press); Cambrian lithostratigraphy after Lendzion (1983a, b); geochronology after Compston *et al*. (1995), Malinowski *et al*. (2005), Krzemińska *et al*. (2007), International Commission on Stratigraphy (2010) brian (e.g. Landing, 1994; Geyer, Uchman, 1995; Paczeŝna, 1996; Landing *et al*., 2007). The lower boundary of the Ediacaran System runs at the base of the Sławatycze or Żuków formations overlying either the crystalline basement (north-eastern region of the Podlasie Depression) or the (?) early Neoproterozoic Polesie Formation (Lublin slope of the East European Craton) along unconformity surfaces (Fig. 5).

In terms of sequence stratigraphy, two 2nd order depositional sequences (A and B) have been identified within the Neoproterozoic sedimentary cover of the Lublin–Podlasie basin (Fig. 6). Sequence A is the lowermost unit of the sedimentary succession in the Lublin–Podlasie basin. It is defined by major angular or erosional unconformities at the base and top, respectively. Along the basal unconformity, clastic deposits of the Polesie Formation are underlain by the Palaeoproterozoic crystalline basement represented by granitoid, syenite, gneiss and gabbro complexes (Malinowski *et al*., 2005). Sequence A is composed of continental and shallow-marine clastics representing probably the early Neoproterozoic Polesie Formation. The late Ediacaran deposits, discussed in this report and represented by various types of both volcanigenic and clastic rocks, are the lower part of the overlying depositional sequence B spanning the late Ediacaran through Middle Cambrian time interval (Paczeŝna, Poprawa, 2005a, b) (Fig. 6).

The Lublin slope of East European Craton was dipping towards the south-west. Thickness of the synrift late Ediacaran–early Cambrian deposits in wells does not exceed 1000 m and continuously increases from the north-east towards the south-west. The maximum thickness of the complete Ediacaran succession (471.9 m), is observed in the Kaplonosy IG 1 borehole located in the centre of the basin. In the most subsiding, southwestern zone of the basin, the succession is over 436.5 m thick in the Łopiennik IG 1 borehole because the drilling was stopped after boring through a small part of the volcanigenic series. This fact suggests that the succession is very thick in this well, probably exceeding 500–700 m. In the Podlasie zone of the basin, thickness of the Ediacaran deposits decreases towards the north-east to reach 100–150 m in the Białowieża region. The gradual south-westward thickness increase is due to increased subsidence in the basin marginal zone.

The deposits discussed in this report are currently hidden under a cover of the Cambrian, Ordovician, Silurian, Upper Palaeozoic, Mesozoic and Cenozoic formations. The cover reaches the thickness of 4000 to 5000 m.

DATA AND RESEARCH METHODS

The fundamentals of all interpretations concerning clastic facies analysis were based on detailed sedimentological and ichnofacies logging of the late Ediacaran to early Cambrian succession in 14 boreholes distributed across the Lublin–Podlasie basin. In the Lublin region of the study area, these were the following wells: Łopiennik IG 1, Terebiń IG 5, Białopole IG 1, Horodło 1, Niwa 1, Święcica 1, Busówno IG 1, Kaplonosy IG 1, Podedwórze IG 2, Parczew IG 10 and Radzyń IG 1. In the Podlasie region, the investigations focused on the Krzyże IG 4, Mielnik IG 1 and Stadniki IG 1 sections (Fig. 2A). The succession was 2459 m in thickness. The total length of 1753 m of cored intervals were described and thoroughly analysed (Tab. 1).

Detailed facies analyses were concentrated on the late Ediacaran Białopole Formation with its lateral equivalent represented by the Siemiatycze Formation, and on the Łopiennik, Włodawa and Mazowsze Formations (Fig. 5). In Podedwórze IG 2, the studied succession has not been pierced because drilling was stopped within the Łopiennik Formation. The upper part of the Włodawa Formation and the whole Mazowsze Formation are lowermost Lower Cambrian.

Sedimentological and ichnological logging enabled the construction of logs at scales of 1:100 and 1:200, which were the basic material for all studies. Some sections of the sequence, in particular those very important for palaeoenvironments interpretations, were analysed at scales of 1:10 and 1:20. Facies and facies associations were identified in all the investigated wells and indicated on the logs. Facies and fa-

Table 1

Thickness of the studied succession and core yield in the investigated boreholes

Borehole	Thickness [m]		Core yield [%]
	analysed succession	cored intervals	
Radzyń IG 1	209	209	100
Kaplonosy IG 1	233	233	100
Podedwórze IG 2	110	110	100
Białopole IG 1	230	230	100
Mielnik IG 1	18	18	100
Stadniki IG 1	4	4	100
Parczew IG 10	244	221	91
Łopiennik IG 1	371	295	79
Krzyże IG 4	81	50	63
Busówno IG 1	300	145	48
Terebiń IG 5	260	116	45
Horodło 1	230	42	18
Niwa 1	228	34	18
Święcica 1	270	43	16

cies associations were interpreted. Extents and distribution of individual sedimentary environments are presented on generalized stratigraphic-facies logs prepared for each well at the scale of 1:1000.

A high-resolution sequence stratigraphic analysis of the late Ediacaran to early Cambrian succession was determined by the existing drill core material for detailed facies analysis. Such investigations were carried out on almost completely cored boreholes drilled by the Polish Geological Institute in the 1960s, trough the 1980s. These are the following wells: Łopiennik IG 1, Białopole IG 1, Terebiń IG 5, Busówno IG 1, Kaplonosy IG 1, Podedwórze IG 2, Radzyń IG 1 and Parczew IG 10 located on the Lublin slope of the East European Craton, and the Krzyże IG 4 and Mielnik IG 1 wells drilled in the Podlasie Depression (Tab. 1). Unfortunately, material from boreholes drilled in the north-eastern and eastern part of the Podlasie Depression (Białowieża region) is currently lacking. It refers to the following wells: Skupowo IG 6, Waśki IG 2, Podborowisko IG 1, Iwanki Rohozy IG 1–4, Rajsk IG 1, Zabłudów IG 1, Kruszyniany IG 1–6, Krynki IG 1, Terespol 1 and Biała Podlaska 1–2.

Much less drill core material has been acquired from wells drilled by the Polish Oil and Gas Company. To identify lithologies and sequence stratigraphic units, the wireline logs, especially gamma ray curves, were analysed in the Niwa 1, Święcica 1 and Horodło 1 wells (Tab. 1). Detailed lithological interpretation of uncored intervals was based on 1:500 scale gamma ray logs. Gamma ray curves were also used to identify coarsening-upwards progradational and fining-upwards retrogradational cycles, as well as the parasequence boundaries. Aggradational cycles, characterised by block-shaped gamma ray curves, have been identified in the Lower Cambrian of some sections. Despite the fact that the gamma ray curves represent a continuous stratigraphical record and reflect trends in grain-size changes, it must be stressed that the interpretations of sedimentary environments and resulting sequence stratigraphic units cannot be considered unambiguous. Data on the facies development acquired as a result of the above-characterised activities were subsequently calibrated to data obtained from direct sedimentological logging of the existing drill cores.

In order to visualize the phases of evolution of the estuarine system a set of combined facies maps and isochore maps was constructed. The facies maps are based on lithological, sedimentological and ichnological data obtained from detailed investigations of cored intervals of borehole sections. These informations are combined with log interpretatation for uncored intervals of the sections. The facies interpretations are superimposed on isochore maps of the six parasequences PS1–PS6. Each parasequence marks the distinct stage of facies development of estuary.

FACIES – TERMINOLOGY AND DESCRIPTION

When determining individual facies types, the classical definition of facies (Gressly, 1838) was used. The facies is understood as a set of characteristics of sedimentary rock, allowing reconstruction of depositional processes and sedimentary environments in terms of both lithology (lithofacies) and organic composition of rocks (biofacies). In the present report, biofacies types are based on the presence of specific trace fossils. Individual facies are jointed in characteristic vertical and lateral facies associations because they represent various types of depositional events that frequently occur together in the same overall depositional environment. If distinct facies is interpreted in terms of the processes which led to the formation of the sediment, facies associations reflect combinations of processes and therefore environments of deposition.

The facies symbols have been adopted from standard lithofacies codes of Miall (1977, 2000). Each facies code consists of two main elements (Tab. 2). The first element of symbols denotes textural features: capital letters for grain size (*G* – conglomerates, *SC* – coarse-grained sandstones, *SF* – fine-grained sandstones, *SM* – medium-grained sandstones, *F* – mudstones and claystones). The second element is a symbol of structural features: sedimentary structures and trace fossils are denoted by a lowercase letter. Types of sedimentary structures and high frequency of trace fossils are described by the following symbols: *m* – massive, structureless deposits, *l* – low-angle planar, large scale cross-bedding, *p* – high-angle planar, large scale cross-bedding, *h* – horizontal stratification or lamination, *r* – ripple cross-lamination, *f* – flaser lamination, *o* – lenticular lamination, *w* – wave lamination, *x* – various types of cross-bedding unidentifiable in drill core, *t* – abundant trace fossils. For example, the symbol *SFlt* denotes fine-grained sandstones with low-angle planar cross-bedding and abundant trace fossils. Two new symbols have also been introduced to accentuate specific facies features of the analysed material: *Fh* – very thinly laminated fine-grained sandstone-mudstone-claystone heterolith and *FS* – thinly bedded fine-grained sandstone-mudstone or sandstone-claystone heterolith (Tab. 2). In both the symbols the letter *F* denotes fine-grained texture of heteroliths, the letter *h* indicates laminated structure of the deposit. The letter *S* in the symbol *FS* denotes sandstone as a stable element of sandstone-mudstone and sandstone-claystone heteroliths.

The terminology and classification of sedimentary structures described in this report require a brief comment because there are no standards in this field. In determining and describing the origin of individual structures, terminology and classification of Zieliński (1998) was used, as it is very clear and coherent. The Zieliński's terminology clearly defines and makes a distinction between cross-lamination and cross-bedding, referring the latter to the structures that developed from migration, lateral accretion and burial of large sea-floor landforms of up to 6.0 cm in height. The terminology also takes into consideration different genetic aspects of the structures. Cross-lami-

Table 2

Facies identified in the latemost Ediacaran and earlymost Lower Cambrian deposits

Facies type (code)	Lithology, lithologic components, sedimentary structures	Trace fossils
1	2	3
Gm	Massive conglomerates, grain-supported to matrix-supported, composed of poorly and well-rounded dark grey quartz, angular pink feldspar or well-rounded claystone and mudstone clasts within sandy matrix. Bed thickness: 0.15–2.0 m	absent
Gl	Conglomerates, grain-supported with low-angle (10–15°) cross-bedding, composed of honey-coloured quartz clasts. Clasts are well rounded, 0.5 to 1.0 cm in diameter. Bed thickness: 0.25–11.0 m	absent
SC	Very coarse-grained sandstones, massive, containing quartz grains, up to 0.5 cm in diameter or quartz and feldspar grains. Bed thickness: 0.2–6.0 m	absent
SCl	Coarse-grained sandstones with low-angle (10–12°) planar cross-bedding. Bed thickness: 0.5–4.0 m	absent
SCh	Coarse-grained sandstones, horizontally stratified. Bed thickness: 0.3–0.4 m	absent
SCp	Coarse-grained sandstones with high-angle large scale cross-bedding (30–40°). Bed thickness: 0.25–6.0 m	absent
SCr	Coarse-grained sandstones with ripple cross-lamination. Bed thickness: 0.4–0.5 m	absent
SCm	Massive coarse-grained sandstones. Bed thickness: 0.4–3.0 m	absent
SM	Massive medium-grained sandstones, rare feldspar and quartz grains 0.3–0.5 cm in diameter. Bed thickness: 0.5–1.0 m	absent
SMl	Medium-grained sandstones with large-scale low-angle planar bedding (10–15°). Bed thickness: 1.0–1.5 m	absent
SMr	Medium-grained sandstones with ripple cross-lamination. Bed thickness: 2.0–4.0 m	absent
SFm	Massive fine-grained sandstones with rare quartz grains 0.4–0.8 cm in diameter. Bed thickness: 0.3–8.0 m	absent
SFl	Fine-grained sandstones with low-angle planar cross-bedding (10–15°), brownish or variegated in colour. Bed thickness: 0.2–2.0 m	absent
SFh	Fine-grained sandstones, horizontally stratified. Rare trace fossils. Bed thickness: 0.5–1.5 m	*Skolithos* isp. *Monocraterion* isp.
SFht	Very fine-grained sandstones with horizontal lamination. Very abundant trace fossils. Bed thickness: 0.5–5.0 m	*Planolites beverleyensis* *Teichichnus rectus* *Neonereites uniserialis*
SFlt	Fine-grained sandstones with low-angle (10–15°) planar cross- bedding. Very abundant trace fossils. Bed thickness: 0.5–4.0 m	*Monocraterion* isp.
SFp	Fine-grained sandstones, high-angle (25– 40°) planar cross-bedding. Bed thickness: 2–9 m	*Monocraterion* isp.
SFrt	Fine-grained sandstones with ripple cross-lamination, lamina sets 0.5–2.0 cm thick. Very abundant trace fossils. Bed thickness: 0.5–4.0 m	*Monocraterion* isp. *Palaeophycus* isp. *Rosselia* isp. *Diplocraterion* isp. *Bergauria major* *Bergauria irregulara* *Teichichnus rectus*
SFr	Fine-grained sandstones with ripple cross-lamination. Bed thickness: 0.25–0.30 m	absent
SFf	Fine-grained sandstones with flaser lamination. Bed thickness: 0.30–4.0 m	absent
SFo	Fine-grained sandstones with lenticular lamination. Bed thickness: 0.15–0.20 m	absent

T a b l e 2 cont.

1	2	3
SFw	Fine-grained sandstones with wavy lamination. Bed thickness: 0.15–0.30 m	absent
Sx	Fine-medium and coarse-grained sandstones with various types of cross-bedding. Bedding types unrecognized in the drill core. Very abundant glauconite. Bed thickness: 0.5–2.0 m	*Planolites montanus* *Monocraterion* isp. escape structures *Planolites beverleyensis*
FS	Thinly bedded fine-grained sandstone–mudstone and sandstone–claystone heterolith, bed thickness from 2 to 15 cm. Abundant flaser lamination in sandstone beds, rare lenticular and horizontal lamination. Abundant trace fossils	*Teichichnus rectus* *Teichichnus* isp. *Planolites montanus* *Bergaueria major* *Palaeophycus* isp. *Skolithos* isp. *Bergaueria* isp. *Diplocraterion* isp. *Treptichnus bifurcus* *Gyrolithes polonicus*
Fh	Very thinly laminated fine-grained sandstone–mudstone–claystone heterolith, lamina thickness below 1 mm, horizontal and flaser lamination, rare lenticular or wavy lamination, numerous reactivation surfaces and 5–25 cm thick fine-grained sandstone interbeds. Numerous small pyrite concretions. Very abundant Cyanophyta *Vendotaenia antiqua* forma *secunda* and *Vendotaenia antiqua* forma *tertia*. Very small diameter of burrows of deposit-feeders, up to 1 mm	*Planolites montanus* *Torrowangea rosei* *Palaeopascichnus delicatu*s *Gordia* isp. *Cochlichnus* isp. *Helminthopsis irregularis* *Bilinichnus simplex*
F	Dark grey, variegated, greenish and cherry massive mudstones and claystones with very rare fine-grained sandstone interbeds (5–10 cm thick) and numerous pyrite layers and concretions. In black claystones very abundant Cyanophyta *Vendotaenia antiqua forma quarta*	absent

nation is formed in the lower part of lower flow regime and represents low-energy structures. Cross-bedding is produced in the upper part of lower flow regime or even in the upper flow regime of high-energy flow. A similar discrimination was made as for horizontal lamination, which forms in the lowest energy conditions, and horizontal bedding which is produced in the high energy flow. In this article, the terminology of horizontally thinly laminated sandstone-mudstone-claystone heteroliths was adopted from Gradziński and Doktor (1996).

DEPOSITIONAL SYSTEMS

A depositional system is an uninterrupted set of depositional process-related sedimentary environments. Active modern and ancient systems are abandoned and buried and become three-dimensional rock bodies (assemblages of lithofacies) with defined areal extent and stratigraphic thickness (after Allen, Allen, 1990; Emery, Myers, 1996; Galloway, Hobday, 1996; Pieńkowski, 2004).

The definition of a depositional system does not involve all aspects of facies analysis, focusing exclusively on lithofacies aspects directly related to the sediment. An essential addition is the introduction of a biological element. To adapt the definition of a depositional system to the requirements of the present report, it has been assumed that biological aspects of facies analysis are represented by trace fossils (Pacześna, 2001) which are the tool of ichnofacies analysis.

Depositional systems were defined based on both facies associations determining their classification and depositional processes which were crucial for their development. The names of individual systems were established based on either the dominant sedimentary environment or assemblages of sedimentary environments. Sedimentary structures and accessory lithologic components (sedimentological analysis) as well as trace fossils (ichnofacies analysis) were the main tools used in determining sedimentary environments.

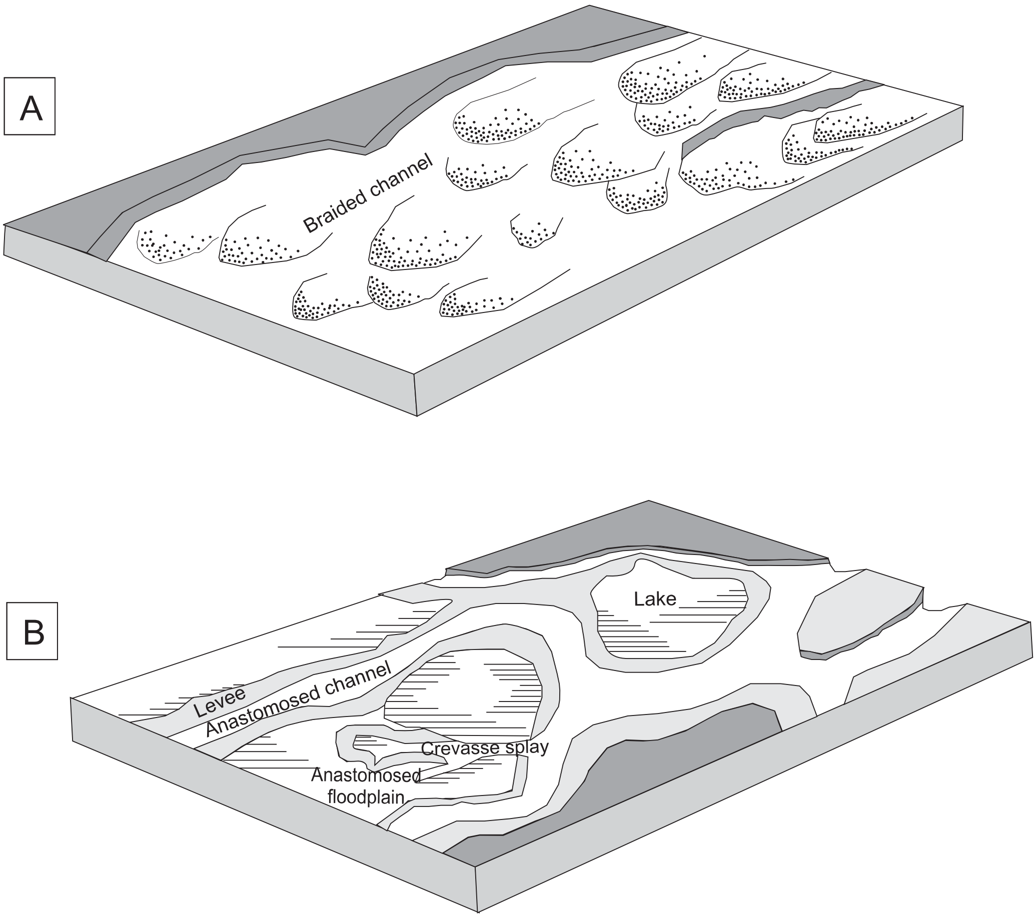

Fig. 7. Braided and anastomosed fluvial style. A – distal to middle reaches sand-dominated braided river system (after Miall, 1996), B – anastomosed system river with low sinusoity, branching channels and floodplain with crevasse splays (after Smith, 1983; Miall, 1996)

Two groups of depositional systems have been identified within the investigated siliciclastic succesion: alluvial, estuarine, and open coast system.

The following depositional systems have been recognized in the group of alluvial systems:

Alluvial fan. An alluvial fan typically is cone-shaped piles of detritus with channels of streams radiating from the apex. The majority of alluvial fans form at the break in slope at the edge of an alluvial plain or at the foot of highlands where streams confined by valley walls emerge into an adjacent lowland. Alluvial fans are composed of two types of sediment: deposit gravity flows and braided stream deposits and are very frequent in arid and semi-arid regions.

Braidplain. A braidplain is a relatively large, flat area with a constant slope, formed by the deposition of sediment over a long period of time by one or more braided rivers (Fig. 7A). The braidplain representing the region over which the braided channels and floodplains have shifted over geological time.

Anastomosed alluvial plain. An anastomosed alluvial plain is a large, plane area, created by the deposition of sediment over a long period of time by one or more anastomosed system rivers (Fig. 7B). To accumulate and preserve anastomosed river deposits low gradient and subsidence of the alluvial plain has to be maintained for considerable geological time.

The range of depositional systems representing sedimentation in the estuary and in its individual segments is given after Dalrymple *et al.* (1992), Einsele (2000) and Dalrymple, Choi (2007). The group of estuarine systems consists of the following depositional systems:

Tidal flat. A tidal flat are smooth, seaward-dipping surfaces dissected by tidal channels and form stretches of shoreline in all tidal settings. They can comprise a large part of estuaries (Fig. 8A, B), lagoon, bays, delta plains, or mangrove swamps. The nature and extent of tidal flats depends on tidal range, wave power, shoreline gradient and the amount and type of available sediment. Tidal flats are most widespread in macrotidal settings (Fig. 8A). Tidal flats can be divided into three parts: supratidal zone above normal high tide level, intertidal zone between high and low tides level and subtidal zone below low tide level. Intertidal zone can be subdivided into three parts from lower to higher elevated areas: sandflat, mixed sand-mudflat and mudflat (Fig. 8A).

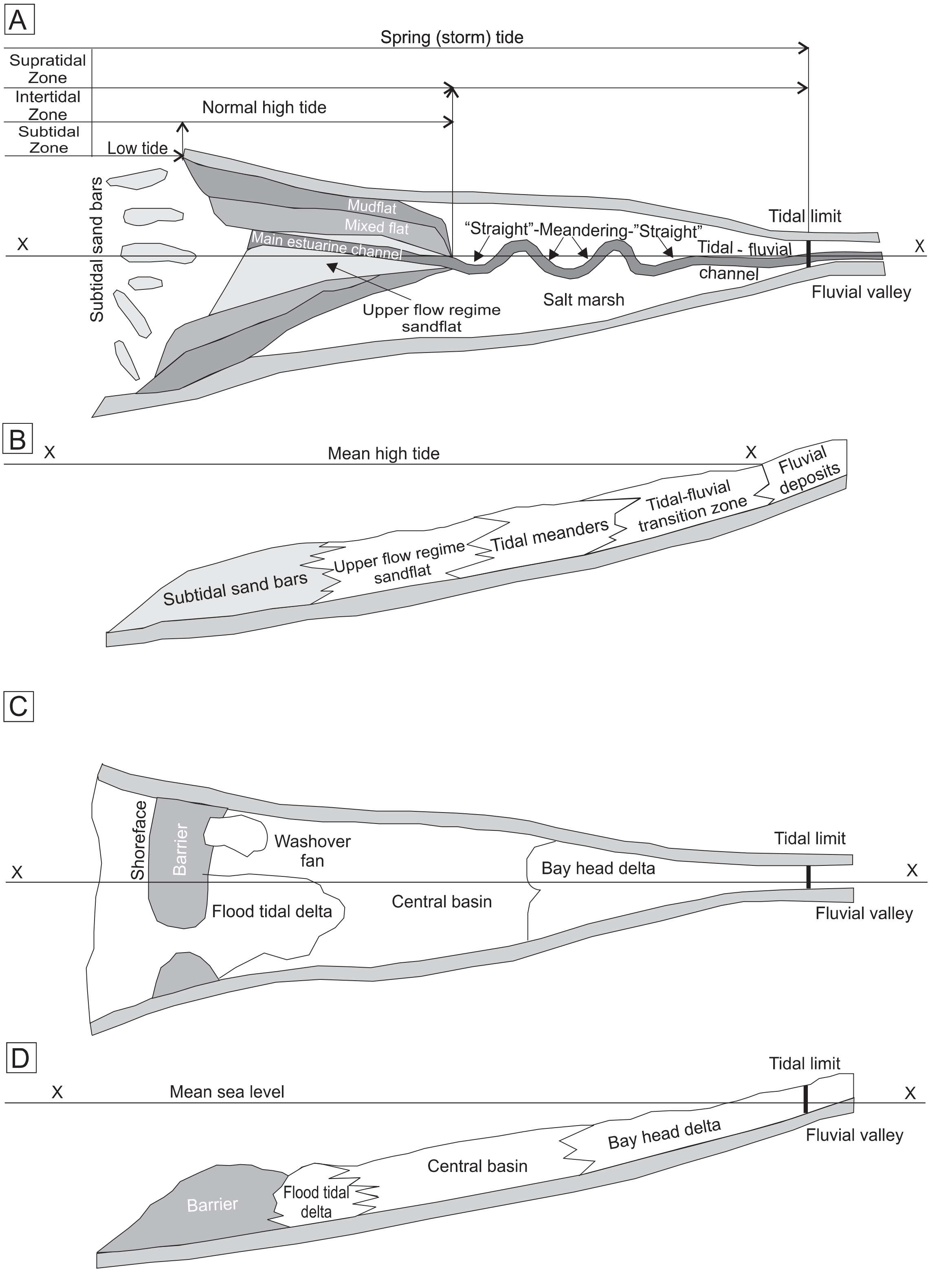

Fig. 8. Schematic models of estuaries. A – morphological elements an idealized tide-dominated estuary in plan view with zonation of tidal environments, B – sedimentary environments in longitudinal section within tide-dominated estuary (after Dalrymple *et al*., 1992; Einsele, 2000, modified), C – morphological components an idealized wave-dominated estuary in plan view, D – sedimentary environments in longitudinal section within wave-dominated estuary (after Dalrymple *et al*., 1992)

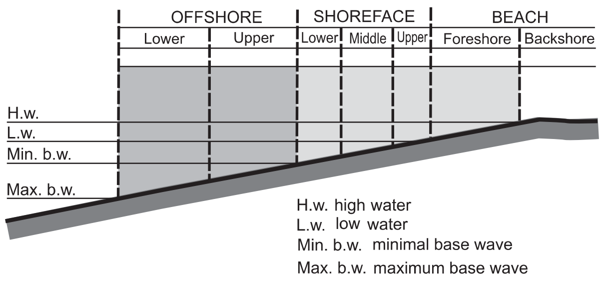

Fig. 9. The distribution of nearshore environmental zones relative to fairweather (normal) wave base (after MacEachern, Pemberton, 1992)

Bay-head delta. A bay-head delta is a complex association of geomorphological settings, sediment types, at the point where a freshwater source enters an estuarine water body. At the mouth of the river channel, the flow velocity is abruptly reduced as the river water enters the standing water a central basin in the wave-dominated estuary (Fig. 8C, D). The river channel mouth is the site of deposition of bedload material.

Central bay and estuary mouth. A central bay is uniform, lower energy environment in the deeper and quieter part of estuary and is formed landward of barrier bar deposits in wave-dominated estuaries (Fig. 8C, D). Central bay typically comprise organic-rich mud and sandy mud. Concentrations of organic material are very high, causing a black to dark appearance in the sediment.

Open coast depositional system. Two main nearshore environmental zones of the open coast have been established by MacEachern, Pemberton (1992): shoreface extending between mean low water and minimum (fairweather) wave base, and offshore encompassing the zone extending between minimum (fairweather) wave base and maximum (storm-weather) wave base. The shoreface is a seaward sloping, sandstone depositional wedge. It is subdivided (from seaward to landward) into a lower, middle and upper shoreface. The lower shoreface is situated between fairweather (normal wave base) and maximum wave base (storm wave base). The middle shoreface extends over the zone of shoaling and breaking waves. The upper shoreface is located in the high energy surf zone. The offshore zone lie below minimum (fairweather) wave base and above maximum (storm-weather) wave base. The offshore are subdividied into two subzones: lower lies directly above maximum wave base and upper extends directly below minimum wave base (Fig. 9).

GROUP OF ALLUVIAL DEPOSITIONAL SYSTEMS

Alluvial fan depositional system

Description. The facies spectrum of alluvial fans was found only in the Parczew IG 10 section. It is dominated by conglomerate facies *Gl* (Fig. 10B). The monomictic conglomerates are composed of poorly rounded quartz clasts (ca. 1–5 cm in diameter) and fragments of granitoid rocks of the crystalline basement. The clasts are embedded in a sandstone matrix. Coarse-grained sandstone facies *SCm* is subordinate; it is represented by thin layers within conglomerate facies *Gl*. The conglomerate bed overlies very coarse-grained sandstones of facies *SCl* containing distinctly developed low-angle cross-bedding.

Interpretation. The angularity of clasts indicate a very short transport of clastic material in the environment of sheetfloods and braided streams periodically flowing across the fan surface. Facies *SCl* and *SCm* are associated with upper flow regime conditions. They reflect strong flows in shallow braided channels that developed on the alluvial fan surface. The alluvial fan setting immediately overlies the crystalline basement. The following facies succession is observed in the lowermost part of section: *SCl*→*SCm*→*Gl*→*SCl*→*Gl* (Fig. 10B). The succession shows coarsening-upward grain-size grading. Coarsening-upward cycles are typical of prograding alluvial fan lobes in their proximal parts (López-Blanco, 1993; Blair, McPherson, 1994; López-Blanco *et al.*, 2000; Rasmussen, 2000).

Alluvial fan depositional system is best developed in the north-eastern region of the Lublin–Podlasie basin near Białowieża. Alluvial fan deposits, were encountered in the Iwanki Rohozy IG 3, Skupowo IG 6, Rajsk IG 1 and Podborowisko IG 1 sections. Unfortunately, the sections are currently unavailable for investigations because the drill cores have been discarded. The conclusions about the origin of the deposits have been inferred from desccriptions given in a number of publications (e.g. Znosko, 1965) and archival materials. The alluvial fan origin is suggested by an increased content of polymictic conglomerates showing neither grain rounding nor clast segregation, and by the presence of cross-bedded coarse-grained sandstones in the deposits. The conglomerates can represent debris flow deposition, whereas the sandstones were deposited from sheetfloods or braided streams flowing across the fan surface (Pacześna, 2006).

Braidplain depositional system

Braided channels facies association

Description. The coarsest grained deposits are reported in the sections located near the source area represented by the crystalline basement elevations. In the north-eastern region of the basin (Krzyże IG 4 section), very coarse-grained arkosic sandstone facies *SC* is represented by 0.2–6.0 m thick sets. Most of the section is represented by packages of medium-grained sandstone facies *SM* and *SMl*. Fine-grained facies are represented by sandstone facies *SFm* and *SFh*. Fine-grained sandstone beds are much thinner as compared with

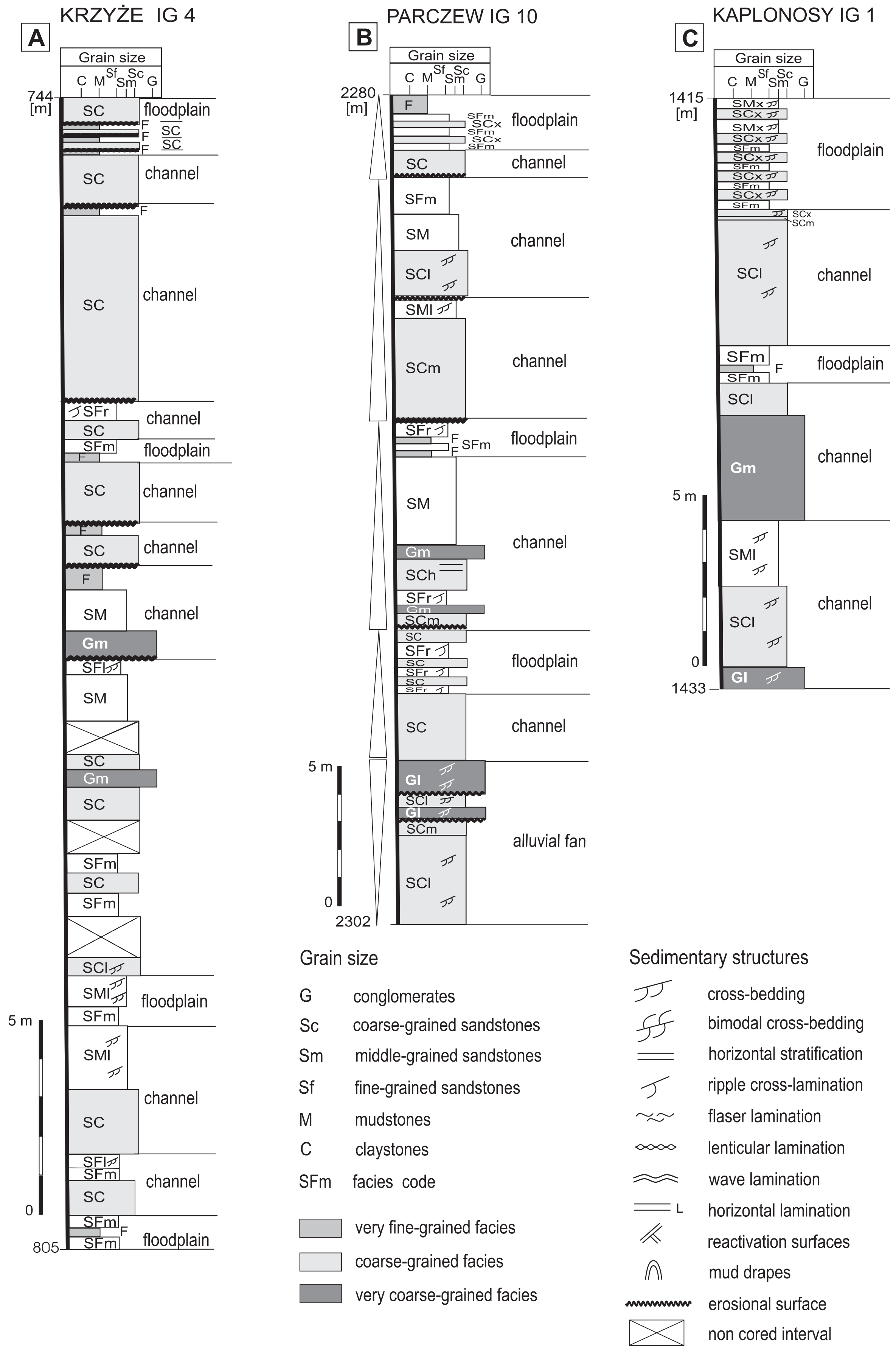

Fig. 10. Facies architecture of coarse-grained braided river successions in the Krzyże IG 4, Parczew IG 10 and Kaplonosy IG 1 borehole sections. Note different types of braided channel fill in the presented sections: single- and multi-storey fills in the Krzyże IG 4 and Parczew IG 10

coarse-grained sandstone beds, ranging from 0.3 m up to 3.0 m. Fine-grained sandstone facies *SFr* is subordinate in the sections, forming 0.3 m thick packages. In the central part of the basin, coarse-grained facies *SCl*, *SCx*, *SCm* occur in the Kaplonosy IG 1. Coarse-grained sandstones and conglomerates form 0.5 to 3.0 m thick packages. Sandstone facies *SMr* and *SMl* occur in 2 to 4 m thick beds. Fine-grained sandstone facies *SFm* are subordinate and represented by 0.2 to 0.5 m thick layers. Another coarse-grained fluvial succession occurs in the Parczew IG 10. It is composed mainly of conglomerate facies *Gl*, and very coarse-grained sandstone facies *SC*, *SCm* and *SCl*. Medium-grained sandstone facies *SM* is subordinate and forms a rare, 2.5 m thick bed. Fine-grained sandstones are represented by facies *SFr* forming numerous very thin packages ranging from 0.2 to 0.3 m in thickness. Sandstone facies *SFm* occurs in a single 1 m thick layers in the upper part of the section. Facies *SCm*, *SCl* and *SMl* are dominant in the middle of the interval and are represented by 0.5 to 1.0 m thick beds. In the cored, lower part of the Busówno IG 1 section, coarse-grained facies *SCl* and *Gm* predominate in 0.5 to 3.0 m thick beds. Fine-grained sandstone facies *Sfm* is subordinate.

Interpretation. The group of sections represents a range of very coarse-grained facies types includes the sections of Krzyże IG 4, Kaplonosy IG 1, Parczew IG 10 and Busówno IG 1, dominated by very coarse-grained arkosic sandstone facies *SC*, *SCm*, *SCl* and *SCx* (Figs. 10, 11, 12C, 13A). Fine-grained facies *SFm* and *SFh* occur in lower proportions. Sediments representing facies *SC*, *SCm* and *SM* were accumulated under upper flow regime from tractional deposition in upper planar bed conditions (Zieliński, 1998; Eriksson *et al*., 2006). Facies *SMl* and *SFh* may represent deposition of low megaripples or antidunes.

An extremely high percentage of very coarse-grained arkosic sandstone facies is observed in the Krzyże IG 4 section. This section is characterised by the occurrence of repeated facies suits, most probably corresponding to successive braided fluvial cycles (Fig. 10A). Cyclicity of depositional processes in braided rivers is less distinctly pronounced as compared with meandering rivers. Braided river cycles are also more often incompletely developed (Miall, 1996; Zieliński, 1998). The following vertical facies successions are observed in the Krzyże IG 4 section:

SC→*SFm* →*SFl* (channel);

SC →*SMl* (channel) →*SFm*→*SMl* (floodplain);

Gm→*SM*→*F* (channel) →*SC*→*F* (channel) →*SC*→*F* (channel) →*SC*→*F* (channel) →*SFm*→*SC*→*SFr* (floodplain).

The two first vertical successions represent a record of mid-channel transverse bars in a shallow braided-river channel under very intense sand accumulation during an abrupt drop of flood water level. Bar deposition occurs when flow decelerates. In general, both the successions are a record of decreasing energy flow in a braided river channel. The third succession represents a transverse bar with a record of declining flow in the channel. The declining flow resulted in deposition of a mud layer (facies *F*). Another interpretation for the origin of mudstone facies *F* is that it was deposited in a stagnant, abandoned inter-bar channel where the mud particles were deposited from suspension (Lindsey, Gaylord, 1992). The average thickness of one-member cycles composed of only coarse-grained channel deposits is 4.1 m. Similar variability of the braided channel-fills has been observed in the late Ediacaran Tanafjord–Varangerfjord sedimentary basin, northern Norway (Røe, Hermansen, 1993).

In the lowermost part of the Krzyże IG 4 section uncomplicated channel fills are predominant, corresponding to a single depositional event resulting in the formation of a single transverse bar (Fig. 10A). Thickness of such channel fills is small, up to 2.5 m. The upper part of the section is composed of four overlying packages of sandstone facies *SC* and *F*. Poorly developed erosional surfaces are observed at the bases of sandstone packages of facies *SC*. Their presence suggests four individual depositional and erosional events. These depositional processes resulted in the formation of four transverse bars, each followed by periods of flow cessation. The above-characterised pattern of the channels indicates that these were multi-storey sandy channel fills (Miall, 1996; Ray, Chakraborty, 2002; Paredes *et al*., 2007) (Fig. 10A). They are typical of large braided rivers (Komatsubara, 2004).

In the Parczew IG 10 section, an alluvial fan succession is overlain by three fluvial cycles (Fig. 10B). The following vertical facies successions have been identified there:

SC (channel) →*SFr*→*SC*→*SFr*→*SC*→*SFr*→*SC* (floodplain);

SCm→*Gm*→*SFr*→*SCh*→*Gm*→*SM* (channel)→*F*→*SFm* →*F*→*SFr*→ (floodplain);

SCm→*SMl* (channel);

SCl→*SM*→*SFm* (channel);

SC (channel) →*SFm*→*SCx*→*SFm*→*SCx*→*F* (floodplain).

The first cycle represents a simple channel fill. The two next cycles are multi-storey channel fills. Each channel is bounded at the base by an erosional surface (Fig. 10B). Average thickness of channel cycles in the Parczew IG 10 section is small, suggesting a channel depth of 3–4 m. Similarly small channel depths are observed in many modern braided rivers. An example of the Sagavanirktok braided river from Alaska indicates channel depths between 2.8 and 3.2 m (Lunt, Bridge, 2004; Lunt *et al*., 2004). Another example comes from the Calamus River, Nebraska, USA, where the depth of the main channel is below 1 m. The depth of the main channel of the South Saskatchewan River, Rocky Mountains, Canada, is 2–5 m (Sambrook Smith *et al*., 2005, 2006).

Fluvial sections of the Krzyże IG 1(Fig. 11B) and Parczew IG 10 contain massive sandstone facies *SC* and *SCm* alternating with planar, low-angle cross-bedded sandstone facies *SCl*, *SMl* or, rarely, *SFh* (Fig. 11A, C, D). It suggests the deposits orginated from hyperconcentrated flood flows and up-

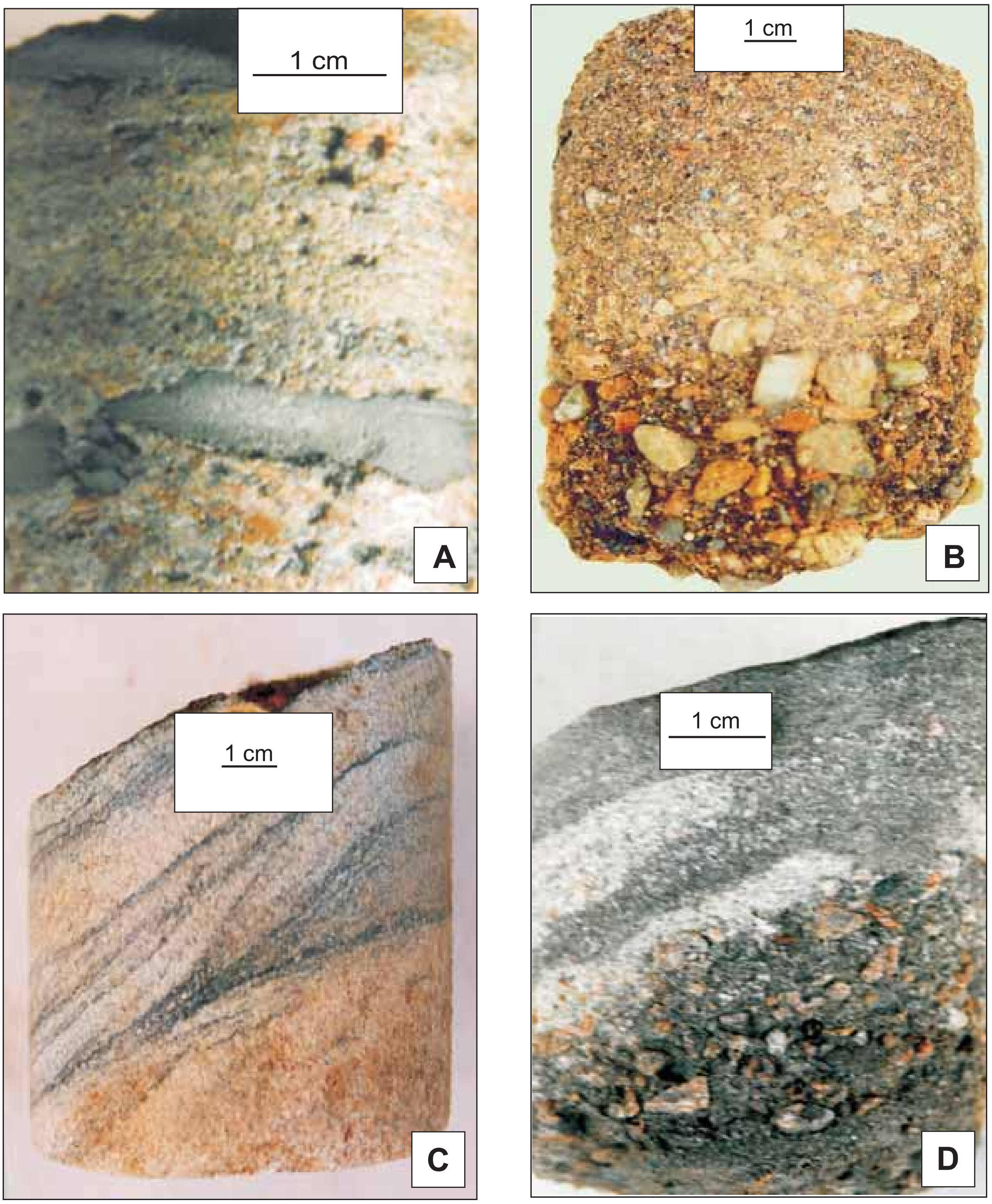

Fig. 11. Channel deposits of braided rivers typical of coarse-grained fluvial successions, late Ediacaran, Siemiatycze Formation

A – very coarse-grained sandstone with single quartz and mudstone clasts distributed along poorly marked cross-bedding, Parczew IG 10 borehole, depth 2290.0 m; **B** – at the base of the core sample a polymictic conglomerate passing up into very coarse arkosic sandstone, Krzyże IG 4 borehole, depth 762 m; **C** – large-scale cross-bedded fine-grained sandstone, Kaplonosy IG 1 borehole, depth 1412 m; **D** – at the base of the core sample a polymictic large-scale cross-bedded conglomerate passing up into large-scale cross-bedded coarse-grained sandstone, Kaplonosy IG 1 borehole, depth 1408.8 m

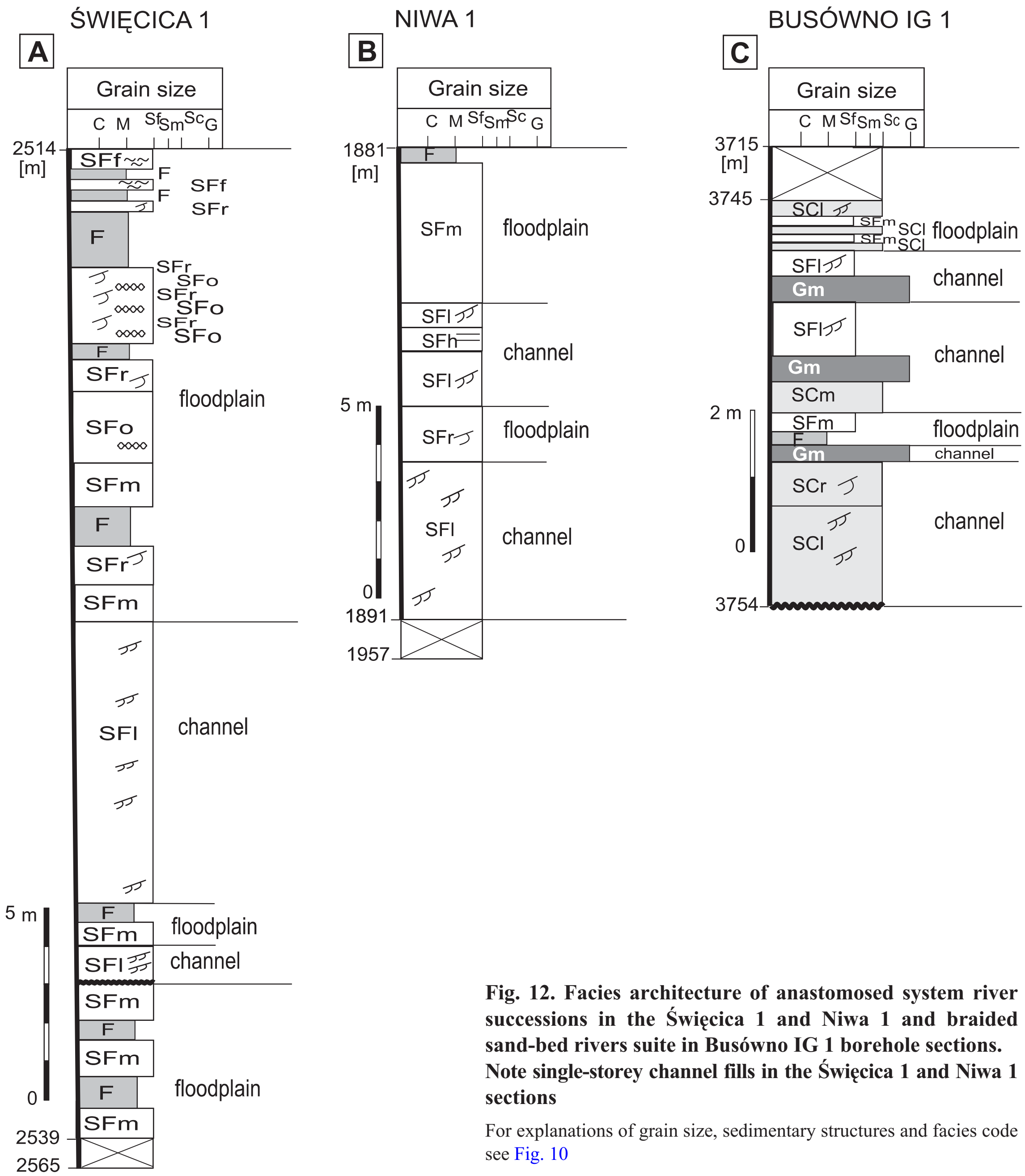

Fig. 12. Facies architecture of anastomosed system river successions in the Święcica 1 and Niwa 1 and braided sand-bed rivers suite in Busówno IG 1 borehole sections. Note single-storey channel fills in the Święcica 1 and Niwa 1 sections

For explanations of grain size, sedimentary structures and facies code see Fig. 10

per regime flow conditions. Both these processes are associated with waning stages of major floods. They are commonly observed in the Precambrian successions of proximal, ephemeral streams and braided rivers flowing down over the surface of alluvial fans (Bhattacharyya, Morad, 1993; Long, 2004; Eriksson *et al.*, 2006). This interpretation is supported by the presence of alluvial fan succession in the lowermost part of the alluvial interval in the Parczew IG 10. In the Krzyże IG 4 section, the occurrence of massive sandstones alternating with planar cross-bedded sandstones could result from deposition in distal parts of an alluvial fan.

Four cycles occur in the lowermost, cored interval of the Busówno IG 1 section (Figs. 12C):

SCl→*SCr* (channel);

Gm→ (channel) *F*→*SFm* (floodplain);

SCm→*Gm*→ (channel) →*SFh* (floodplain);

Gm→*SFl* (channel) →*SCl*→*SFm*→*SCl*→*SFm*→*SCl* (floodplain).

The occurrence of simple cycles indicates waning flow energy in the shallow channels filled with conglomerates and coarse-grained sandstones (Fig. 13A). The average thickness of coarse-grained channel deposits of the cycles is below 2 m.

Four cycles have been identified in the Kaplonosy IG 1 borehole section drilled in the centre of the basin (Fig. 10C).

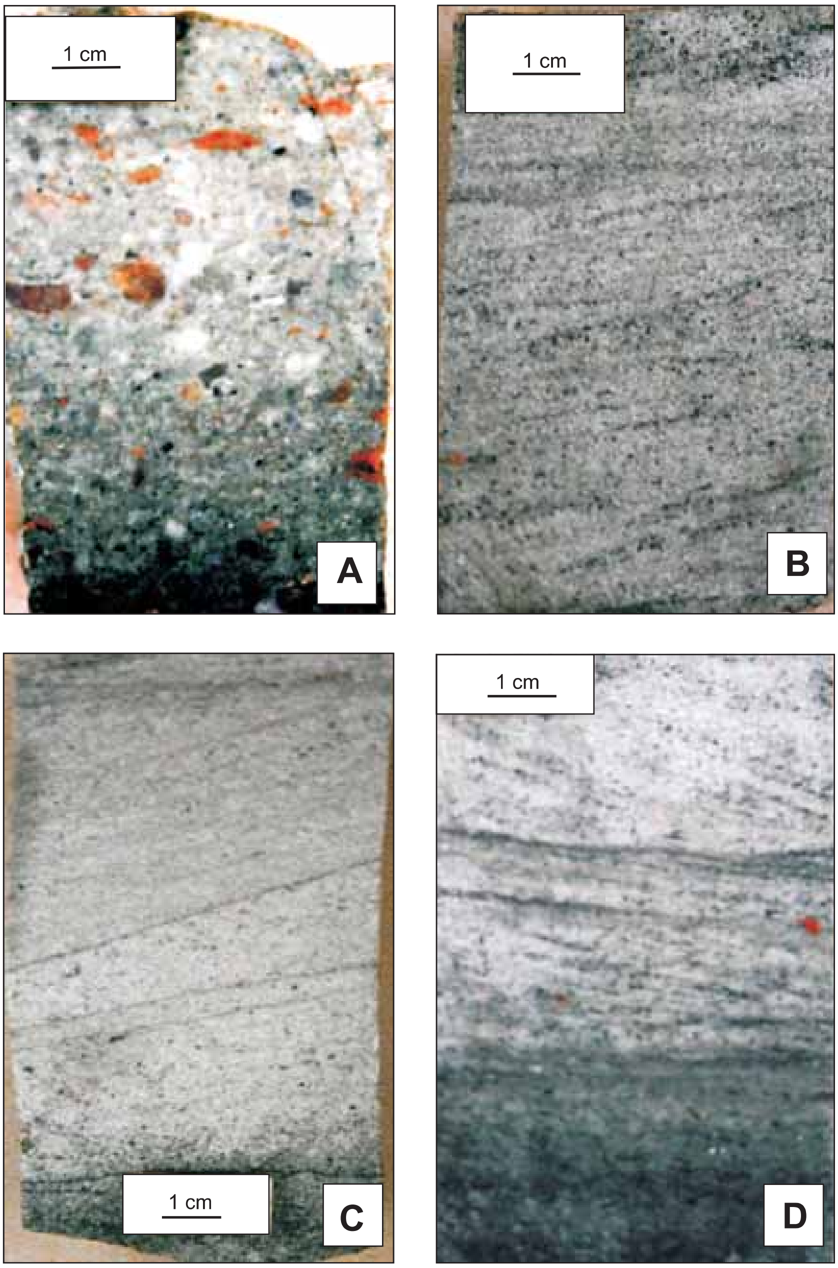

Fig. 13. Braided channel and floodplain deposits typical of distal braided sand-bed river successions, late Ediacaran, Siemiatycze Formation, Busówno IG 1 borehole

A – massive polymictic conglomerate, braided channel facies association, depth 3745.0 m; **B–D** – large-scale low-angle cross-bedded fine-grained sandstone, depth: 3750.0 m (B), 3750.9 m (C), 3752.0 m (D), braided floodplain facies association

Each of the cycle is composed of the following facies succession (from the base to the top):

Gl→*SCl*→*SMl* (channel);

Gm→*SCl* (channel) →*SFm*→*F*→*SFm* (floodplain);

SCl→*SCm*→*SCx* (channel) →*SFm*→*SCx*→*SFm*→*SCx* →*SFm*→*SCx*→*Sfm*→*SCx*→*SMx*→*SCx*→*SMx* (floodplain).

Channel members of the first two cycles are composed completely of high-energy facies deposited in supercritical flow conditions. Channel member of the third cycle is characterised initially by similar, high-energy hydrodynamic conditions. It is suggested by the presence of coarsed-grained sandstone facies *SCl* and *SCm*. The appearance of *SCx* facies in the uppermost part of the channel member probably resulted from decreasing flow energy when the flood surge recedes. The average thickness of the channel part of the cycles is 4.5 m.

No facies packages typical of inter-bar channels are observed in the above discussed fluvial sections. This fact was determined by the type of geological material available for research. Inter-bar channels are characterised by trough cross-bedded cosets. Another typical feature of the depositional environment is a sequence of trough cross-bedded sandstone facies grading upwards into sandstones facies containing horizontal stratification or ripple cross lamination (Zieliński, 1998). The presence of trough cross-bedded sets is very difficult or even impossible for identification in investigated drill cores due to a limited observation field and, commonly, poor condition of drill cores.

Channel deposits bear features of typical distal to middle reaches sand-bed braided channels (Fig. 7A). Their braided nature is confirmed by diagnostic features, including commonness of index coarse-grained facies represented by very coarse-grained sandstones and conglomerates. Another important diagnostic feature is the occurrence of planar low-angle cross-bedding. Planar low-angle cross-bedding is an indicator of transverse bars in shallow braided-river channels. The average thickness of sandstone facies *SCl* and *SMl* varies from 1.5 to 1.6 m. It indicates a relatively small depth of the channels. Transverse bars are depositional megaforms typical of braided sand-bed rivers with very strong aggradation of the river bed (Zieliński, 1993; Miall, 1996). The presence of massive coarse-grained sandstone facies *SC*, *SM* and *SCm* representing supercritical flow conditions suggests deposition in high-energy braided sand-bed rivers with strong aggradation of the river bed.

The prominent factor for stimulating the braided river system development was the late Ediacaran climate favouring chemical and physical weathering. As a result of strong erosional processes, loose detrital material accumulated in alluvial drainage basins. Lack of plant cover favoured mobilisation of material eroded on topographic highs. All these factors resulted in a continuous supply of coarse-grained material to the braided rivers and streams, which built transverse bars of braided channels.

Braided floodplain facies association

Description. The braided floodplain association is represented by a relatively narrow range of facies types. The most common are facies *SFm* and *F*. There is a dependence of quantitative contribution of coarse- and fine-grained facies in individual cycles upon location within the sedimentary basin. Those sections, where braided channels are composed of very coarse-grained deposits, typically contain coarser-grained sediments accumulated on a braided floodplain and commonly representing facies *SC*, *SCx* and *SCm*. They are interbedded with fine-grained sandstone facies *SFm*, *SFr*, *SFf* or *F*. The braided floodplain deposits are commonly thin, ranging from 1.0 to 3.5 m.

Interpretation. The frequent facies *SC*, *SCx* and relative high frequency of ripple facies of the braided floodplain suggest influence of active channel depositional processes onto its area at times of very large flood surges. The interbedding of cross-bedded and massive coarse-grained sandstones with massive fine-grained sandstones or ripple cross-laminated sandstones seems to confirm the occurrence of the process (Zieliński, 1998; Allen, Fielding, 2007). Such process is suggested by the presence of the following facies suites:

SC→ (channel) →*F*→*SC*→*F*→*SC*→*F*→*SC*→ (floodplain) in the Krzyże IG 1 section (Fig. 14C);

SCl (channel) →*SFm*→*SCx*→*SFm*→*SCx*→*SFm*→*Scx* →*SFm*→*SCx*→*SMx*→*SCx*→*SMx* (floodplain) in the Kaplonosy IG 1 section (Fig. 14A);

Gm→*SFl* (channel) →*SCl*→*SFm*→*SCl*→*SFm*→*SCl* (floodplain) in the Busówno IG 1 section (Figs. 12C, 13B–D);

SC (channel) →*SFr*→*SC*→*SFr*→*SC*→*SFr*→*SC*→*SCm* (floodplain) in the Parczew IG 10 section (Fig. 14B).

The interbedding of relatively thin layers of fine-grained sandstone facies *SFm* and *SFr* with mudstone facies *F*, centimetres to decimetres thick is observed in Parczew IG 10 and is characteristic of distal parts of braided floodplains. It is observed, for example, in the late Permian braided systems of Queensland, Australia (Allen, Fielding, 2007) and the lowermost Lower Cambrian braided floodplains of the Mojave Desert, California (Fedo, Cooper, 1990).

All the above discussed braided floodplain successions are typical for a distal floodplain. It should be stressed that braided floodplains of the rivers were not extensive. Compared with a few hundred metres wide river channels, the braided floodplains accounted for merely several percent of the width of the entire braided fluvial system. Contribution of the thickness of very fine-grained facies, especially mudstones and claystones, to the total thickness of floodplain deposits was also small. They accounted for 0% to 20% of all fine-grained components of the floodplain, represented mostly by fine-grained sandstone facies *SFm*. These percentages od mudstone–claystone deposits in the floodplain successions are one of the indicators confirming the braided nature of the late Ediacaran rivers in the basin.

Anastomosed alluvial plain depositional system

Anastomosed river channel facies association

Description. In the south of the central part of the Lublin–Podlasie basin, a group of fluvial sections dominated with fine-grained sandstone facies have been distinguished in the Święcica 1 and Niwa 1 sections (Fig. 12A, B). Fine-grained deposits occur also in the Radzyń IG 1 and Mielnik IG 1 (Fig. 15A, B). Very fine-grained deposits, represented mainly by mudstones and rarely claystones of facies *F* occur in numerous 0.2 to 1.0 m thick beds. Most of the sections is represented by packages of fine-grained sandstone facies *SFl* and *SFm*. Sandstone facies *SFl* forming 0.5 to 6.0 m thick packages. Sandstone facies *SFm* occurs in 0.5 to 3.5 m thick layers. Facies *SFr*, *SFf* and *SFo* are rare, forming packages 0.3 to 2.0 m thick packages. Fine-grained sandstone facies SFh are subordinate and represented by 0.5 to 1.0 m thick layers. Sandstone facies *SM* and *SMl* occur in 2 to 4 m thick beds. Medium-grained sandstone facies *SM* forms 0.5 to 2.5 m thick bed. In the Radzyń IG 1 facies *SCm* and *SCl* occur in the middle of the interval and are represented by 1.0 to

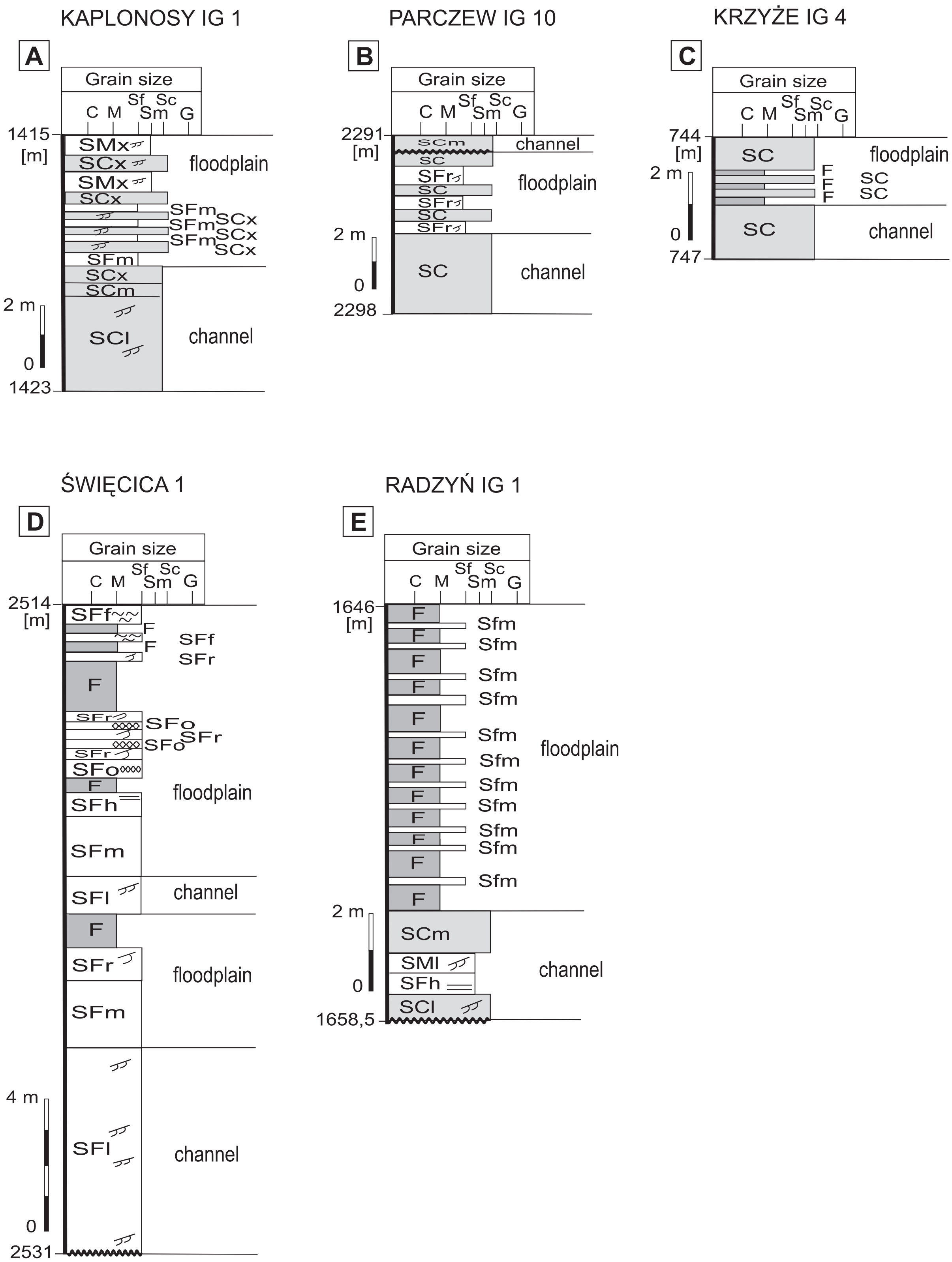

Fig. 14. Facies successions of braided and anastomosed system river floodplains. A–C – braided floodplain, D, E – anastomosed river floodplain

For explanations of grain size, sedimentary structures and facies code see Fig. 10

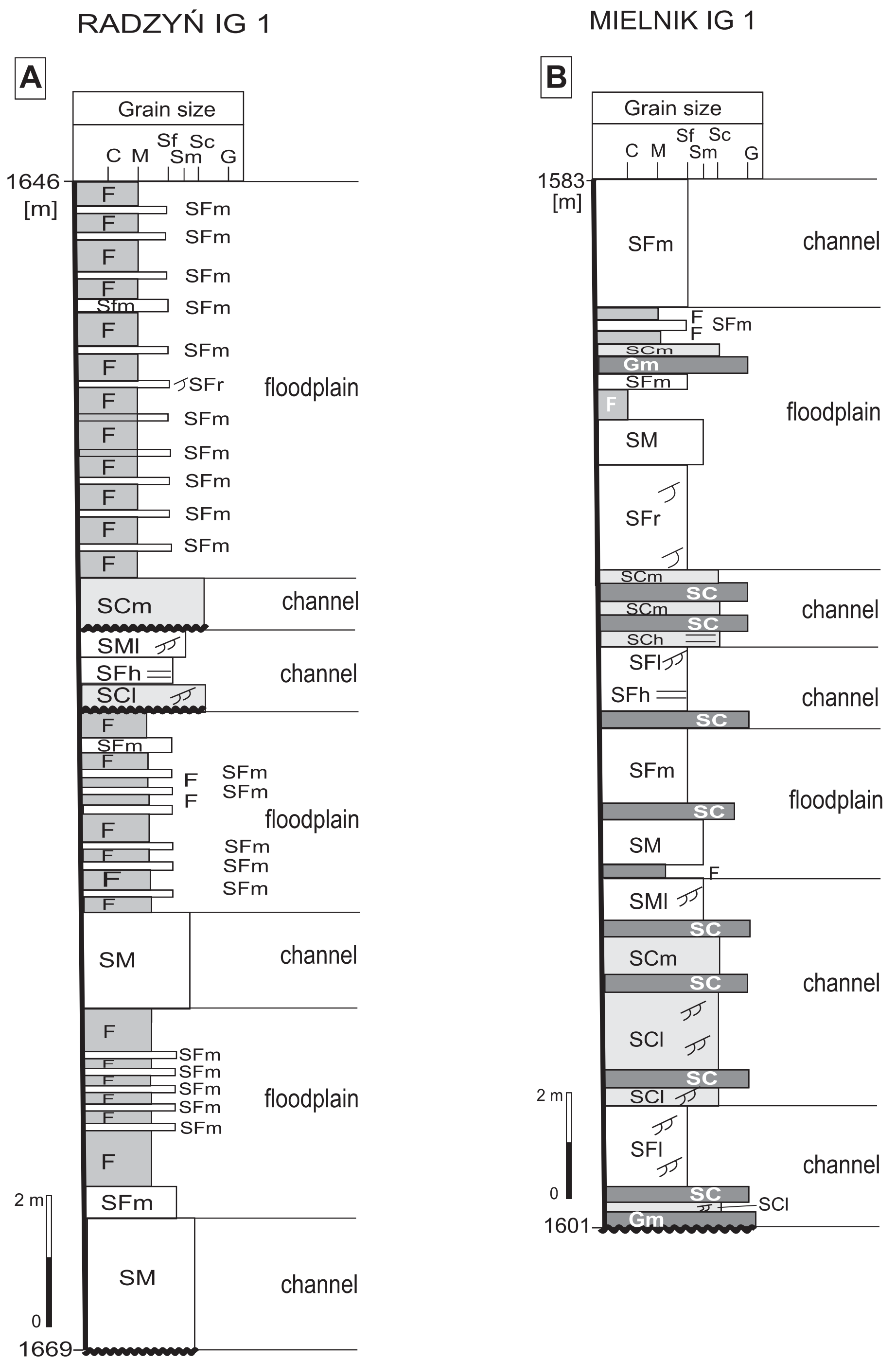

Fig. 15. Facies architecture of anastomosed system river successions in the Radzyń IG 1 and Mielnik IG 1 borehole sections. Note single-storey channel fills in the both presented sections

For explanations of grain size, sedimentary structures and facies code see Fig. 10

0.5 m thick beds. In the Mielnik IG 1 coarse-grained facies *SCl* and *SCm* form 0.2 to 1.5 m thick packages. Conglomerate facies *Gm* forms 0.2 m thick layers.

Interpretation. No coarse-grained facies is observed in cored intervals of the Święcica 1 and Niwa 1 sections (Fig. 12A, B). The lower parts of the Niwa 1 and Święcica 1 sections probably contain coarse-grained sandstone or conglomerate beds, as evidenced by gamma ray logs. There is also a fining-upward trend observed within the fluvial cycles. The cored interval of the Niwa 1 section reveals the presence of three cycles. The following facies successions have been identified, from the bottom to the top (Fig. 12B):

SFl (channel) →*SFr* (floodplain);

SFl→*SFh*→*SFl* (channel) →*SFm*→*F* (floodplain).

Sandstones of facies *SFl*, occurring in two successive cycles, were deposited in a transitional conditions from upper to lower flow regime within low channel bars. Transitional hydrodynamic conditions were relatively long during the first cycle, as evidenced by the large thickness (3.5 m) of sandstone facies *SFl*. The second cycle's facies *SFh* indicates a short episode of supercritical flow during a flood surge. Horizontal bedding forms in fine-grained sands at a depth of 0.25–0.50 m and flow velocity of 1 m/s (Miall, 1996). Transition of facies *SFh* into the overlying facies *SFl* reflects a decrease in flow energy after the flood surge recedes.

In the Święcica 1 section, the base of fluvial interval has not been cored. Only its upper part is available for direct observation. The lower part of the fluvial interval is probably represented by the upper part of the cycle with a facies succession of *SFm*→*F*→*SFm*→*F*→*SFm* (floodplain), and the overlying cycles (Fig. 12A):

SFl (channel) →*SFm*→*F* (floodplain);

SFl (channel) →*SFm*→*SFr*→*F*→*SFm*→*SFo*→*SFr*→*F* →*SFo*→*SFr*→*SFo*→*SFr*→*SFo*→*SFr*→*SFr*→*F*→*SFf*→*F* →*SFf* (floodplain).

Channel members of the first and second cycles are composed exclusively of fine-grained sandstones of facies *SFl*. The presence of low-angle planar cross bedding suggests that they were deposited in the upper part of lower flow regime, within large bedforms of megaripples or transverse bars. The exclusive presence of facies *SFl* is associated with river-bed aggradation during long-term conditions of decreasing flow capacity. The base of channel member of the first cycle is accentuated by a flat 5th order bounding surface without any conglomeratic channel lag layer. Such surfaces bound at the base sand bodies that develop during migration and filling of single river channels (Miall, 1991, 1996). No erosional surfaces are observed at the top of both the channels. Thicknesses of the channels ranges from1 to 6 m. No grain-size fining is observed in these two channel members.

In the Mielnik IG 1 section the following facies successions have been identified from the bottom to the top (Fig. 15B):

Gm→*SCl*→*SC*→*SFl* (channel);

SCl→*SC*→*SCl*→*SC*→*SCm*→*SC*→*SMl* (channel) →*F*→ *SM*→*SC*→*SFm* (floodplain);

SC→*SFh*→*SFl* (channel);
SCh→*SC*→*SCm*→*SC*→*SCm* (channel) →*SFr*→*SM*→*F* →*SFm*→*GM*→*SCm*→*F*→*SFm*→*F* (floodplain);

SFm→ (channel).

Channel member of the first cycle begins with facies *Gm* represented by channel lag conglomerates (Fig. 16D). They were deposited in high-energy conditions and under supercritical flow. They are underlain by a 5th order bounding surface indicating the channel base (Miall, 1996). The channel's facies succession is a record of a change from transitional hydrodynamic conditions of the upper part of lower flow regime represented by facies *SCl* to extremely high-energy conditions of hyperconcentrated flow indicated by the massive coarse-grained sandstones of facies *SC*. The transition *Gm*→*SCl*→*SC*→*SFl* is a record of a single flood surge followed by decreasing flow energy as a result of surge recession. The presence of sandstone facies *SCl* and *SFl* (Fig. 16A, C) suggests strong river-bed aggradation in the zones of decreasing flow capacity. A fining-upward trend is observed within the channel. The total thickness of the channel deposits is 2.0 m. Proportion of fine-grained deposits of facies *SFl* is significantly higher, and their thickness is 1.5 m.

The characteristic succession of twice-repeated very high-energy conglomerate facies *SC* (Fig. 16B) and *SCm* in the second cycle's channel member, separated by a period of deposition of facies *SCl* and *SMl*, is a record of two flood surges followed by periods of decreasing flow capacity, respectively. There was rapid river-bed aggradation in the channel. The channel-fill deposits are 3.3 m in thickness.

The third cycle's river channel was filled mainly with fine-grained deposits under supercritical flow during a flood surge. The presence of facies *SFl* at the channel top is due to surge recession. There was rapid river-bed aggradation in the channel. The facies distribution indicates a fining-upwards trend within the channel. The channel-fill deposits are 1.7 m thick, including 1.5 m thick deposits of fine-grained facies *SFh* and *SFl*.

Facies succession of the fourth cycle suggests very rapid river-bed aggradation under supercritical flows transitioning into hyperconcentrated flows. Thickness of the channel member is 1.2 m. No fining upwards in grain size is observed.

The currently observed fine-grained deposits of facies *SFm* probably represent the lower part of the fifth fluvial cycle. They are overlain above an erosional unconformity by Lower Cambrian deposits. The upper part of the cycle was probably removed during the Lower Cambrian marine transgression. The massive fine-grained sandstones of facies *SFm* were deposited under extremely high hyperconcentrated flow conditions. Thickness of the river-channel deposits is 2 m.

The Radzyń IG 1 section is represented by four cycles (Fig. 15A):

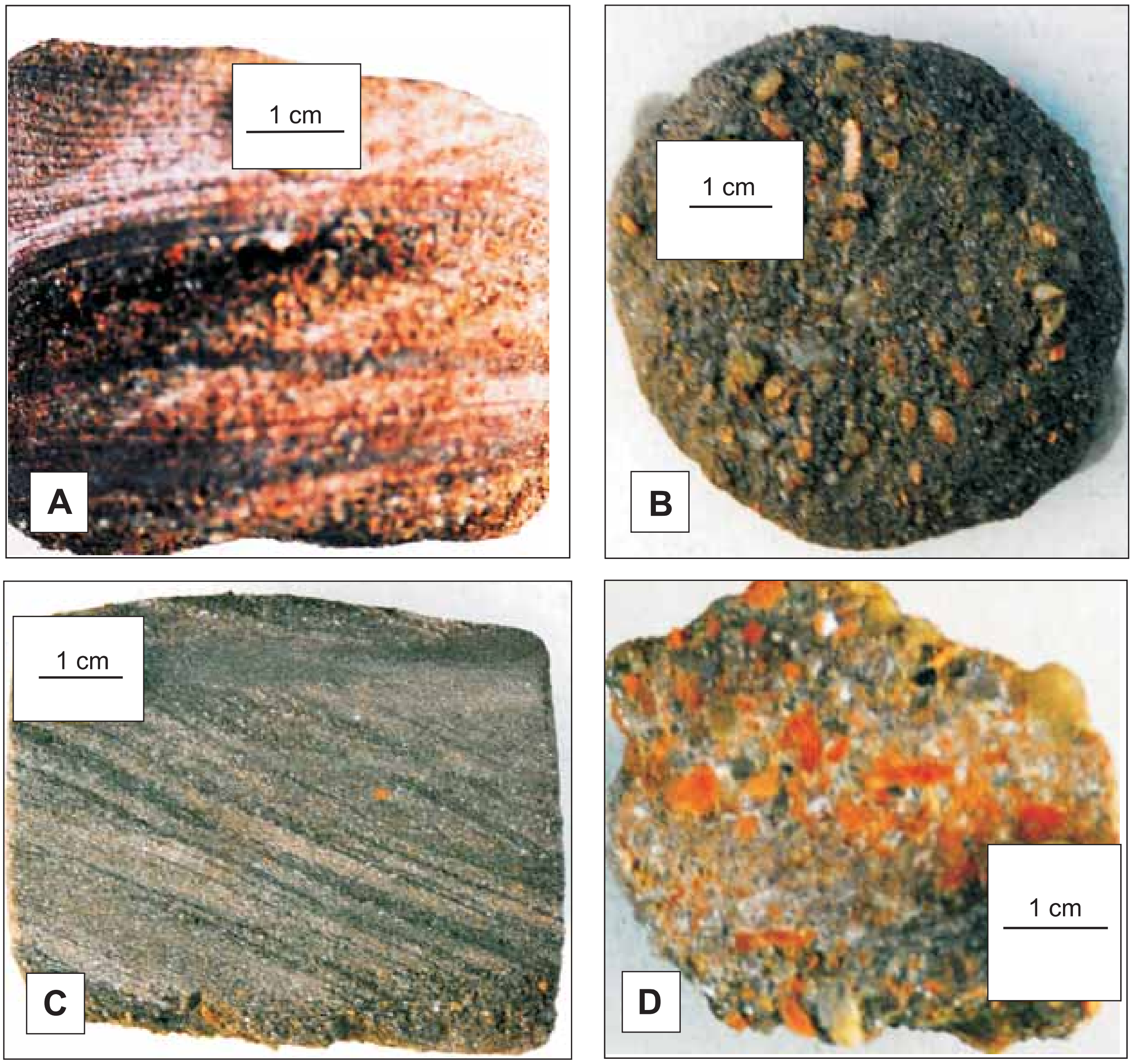

Fig. 16. Coarse- and fine-grained deposits of fluvial succession of Mielnik IG 1 borehole section. Anastomosed river channel facies association, late Ediacaran, Siemiatycze Formation

A – low-angle horizontally bedded coarse-grained sandstones, depth 1600.5 m; **B** – conglomerate with rare large feldspar and quartz grains, depth 1592.0 m; **C** – low-angle cross-bedded fine-grained sandstone, conglomerate with quartz and feldspar grains at the base of drill core sample, depth 1600.0 m; **D** – conglomerate with rare large quartz grains, depth 1600.2 m

SM (channel) → a package of six layers of sandstone facies *SFm* (0.1–0.5 m thick) interbedded with seven layers of mudstone facies *F* (0.1–0.6 m thick) (floodplain);

SM (channel) → a package of seven layers of sandstone facies *SFm* (0.1–0.5 m thick) interbedded with eight layers of mudstone facies *F* (0.1–0.5 m thick) (floodplain);

SCl→*SFh*→*SMl* (channel);

SCm (channel) → a package of eleven layers of sandstone facies *SFm* (0.1–0.5 m thick) interbedded with tvelve layers of mudstone facies *F* (0.1–0.2 m thick) (floodplain).

Channel members of the first and second cycle are composed exclusively of massive medium-grained sandstones. They were deposited during a single depositional event under extremely high-energy conditions, indicating strong river-bed aggradation. Thicknesses of the channel fills of the first (lower) and second (upper) cycles are 2.5 and 1.2 m, respectively. The third cycle is a record of a short event of abrupt increase in hydrodynamic conditions in the rver channel. It is suggested by a transition of facies *SCl* into *SFh*, which is a record of water surge. After the surge maximum, proved by the presence of facies *SFh*, there was a decrease in flow capacity recorded in the underlying facies *SMl*. Thickness of the river-channel member is 1.5 m, of which fine- and medium-grained facies account for a thickness of 1 m. River-channel member of the fourth (uppermost) cycle shows a simple structure. It is composed exclusively of massive coarse-grained sandstones of facies *SCm*. The river channel was characterised by very high-energy flow conditions and rapid river-bed aggradation. Thickness of the river-channel fill is 1 m. High-relief erosional surfaces are observed at the bases of the first, third and fourth cycle. They bound sand bodies migrating along the river bed. No conglomeratic channel lag is observed above. These are the 5th order bounding surfaces according to the classification of Miall (1996).

Sedimentation of the above discussed river-channel deposits occurred in sand-bed river channels probably representing an anastomosed system rivers. The diagnostic features of this river type include:

1. Lack of fining upwards in grain size within the channel succession due to vertical sediment accretion. This feature is particularly well expressed in the Radzyń IG 1 (Fig. 15A) and Święcica 1 (Fig. 12A) sections. Vertical accretion is a typical depositional process occurring in anastomosed river channels (Nadon, 1994; Miall, 1996). Lateral and concentric accretion is rarer (Kirschbaum, McCabe, 1992). Similar anastomosed channels with strong vertical aggradation have been reported by Waksmundzka (2005, 2008) from the Carboniferous succession of the Lublin region (south-eastern Poland).

2. Most of channel fills are composed of fine-grained sandstones.

3. Simple structure of channels composed chiefly of single, large bedforms that developed during a single depositional episode.

4. Lack of ripple-laminated fine-grained sandstones *SFr*, *SFf*, *Sfo* and *SFw* at the top of channel deposits. It indicates both stable hydrodynamic conditions within the channels without decreasing flow energy and lateral stability of the channels.

5. Very high frequency of massive facies represented by sandstones *SFm*, *SM* and *SCm*, and low-angle planar cross-bedded sandstones of facies *SFl*, *SMl* and *SCl*. All these facies types suggest strong river-bed aggradation.

6. Very small thicknesses of river-channel fills as compared with thicknesses of floodplain deposits. The examples of such relation are the Radzyń IG 1 (Fig. 15A) and Święcica 1 (Fig. 12A) sections where the thickness of floodplain deposits in each cycle, corrected for decompaction (Baldwin, Butler 1985) of mudstones and claystones, varies from nearly 70% to 94 % of the total cycle thickness. For instance: the total original thickness of the uppermost cycle in Radzyń IG 1 was approximately 19 m, of which floodplain deposits were almost 18 m thick and accounted for 94% of the total cycle thickness.

Anastomosed rivers form interconnected networks of low gradient channels of variable sinuosity (Fig. 7B). Changes in channel course occur through avulsion result in progressive crevassing of the channel banks. Anastomosed channel pattern preferentially developed in times of rapidly rising base level in subsiding areas. Increase of accomodation space leads to a fluvial succession dominated by vertical accretion deposits (e.g. Smith, 1983; Tornqvist, 1993; Miall, 1996).

Anastomosed river floodplain facies association

Description. The anastomosed river floodplain association is represented by a relatively wide range of fine-grained facies types. In the Radzyń IG 1, Niwa 1 and Święcica 1, only fine-grained deposits are observed. In the Mielnik IG 1 section, fine-grained facies *SFm* and *F* are accompanied by coarse-grained facies *Sc* and medium-grained facies *SM*. Sandstone facies *SFm* occurs in 0.1 to 3.5 m thick beds. In the Radzyń IG 1 section, sandstones of fine-grained facies *SFr*, *SFo* and *SFf* are frequent, forming 0.1 to 2.0 m thick packages. Very fine-grained deposits, represented mainly by mudstones and rarely claystones of facies *F* occur in numerous 0.2 to 1.0 m thick beds. Very rare sandstone facies *SM* occur in 0.8 m thick beds. The floodplain deposits are commonly thick, ranging from 3.0 m in Mielnik IG 1 to 6.5 m in Radzyń IG 1.

Interpretation. Floodplain deposits overlie river-channel sediments, contrasting in facies appearance. An example is the facies successions from the Święcica 1 section:

SFl (channel) →*SFm*→*F* (floodplain);

SFl (channel) →*SFm*→*SFr*→*F*→*SFm*→*SFo*→*SFr*→*F* →*SFo*→*SFr*→*SFo*→*SFr*→*SFo*→*SFr*→*SFr*→*F*→*SFf*→ *F*→*SFf* (floodplain) (Fig. 12A, 14D).

In the facies succession of the first cycle, the transition of high-energy facies *SFm* into very low-energy mudstone facies *F* was probably associated with levee breaks and overflow of water onto the floodplain situated lower in elevation. Sandstone facies *SFm* was deposited under very high-energy flow within crevasse splays during the initial stage of its formation. Mudstone facies *F* was deposited under the lowermost part of lower flow regime during flow deceleration and deposition from suspension.

Compared with the floodplain member of the first cycle, the second cycle's member is much more complicated. It is a record of two episodes of water overflow from the channel onto the floodplain and initial formation of sand crevasse splays of facies *SFm* under supercritical flow conditions. The facies succession *SFm*→*SFr*→*F* is associated with a gradual decline of the flow in the crevasse channel during the formation of ripples by rhythmical bottom transport and deposition from suspension. Due to these processes, the crevasse channel was filled with mudstone facies *F*. A similar facies succession occurs above the lowermost part of the floodplain succession:

SFm→*SFo*→*SFr*→*F*→*SFo*→*SFr*→*SFo*→*SFr*→*SFo*→ *SFr*→*SFr*→*F*→*SFf*→*F*→*SFf*.

After the surge receded, three successive episodes of deposition from rhythmical bottom transport (facies *SFr*, *Sfo* and *SFf*) and deposition of mud from suspension (facies *F*) occurred in the crevasse channel (facies *SFm*). Thickness of the ripple sandstone facies is relatively large and varies from 1 to 2 m. Thickness of the mudstone fills of crevasse splays ranges between 1 to 2 m. The total thickness of the floodplain deposits is about 12 m (not corrected for decompaction) and is twice as much as the thickness of the cycle's river-channel member.

In the cored interval of the Niwa 1 section, which is only a small fraction of the whole fluvial succession, floodplain deposits are fully represented only in the second cycle. The facies succession *SFm→F* is a record of high-energy deposition in a crevasse splay channel, followed by flow decline and mud deposition from suspension. The large thickness of crevasse splay deposits (3.5 m) suggests a prolonged deposition under supercritical flow conditions during water surge within the crevasse splay channel, and strong channel-bed aggradation.

Another group of deposits is represented by fine-grained sediments forming characteristic successions of rhythmically alternating sandstones of facies *SFm* and mudstones of facies *F* overlying the channel members in most completely developed successions from the Radzyń IG 1 and Mielnik IG 1 sections. These are the following floodplain suites in the Radzyń IG 1 section (Figs. 14E, 15A):

– a package of six layers of sandstone facies *SFm* (0.1–0.4 m thick) interbedded with six layers of mudstone facies *F* (0.2–0.9 m thick) in lowermost cycle;

– a succession of seven layers of sandstone facies *SFm* (0.1–0.2 m thick) interbedded with eight layers of mudstone facies *F* (0.1–0.5 m thick) in the second cycle;

– a suite of eleven layers of sandstone facies *SFm* (0.1–0.2 m thick) interbedded with tvelve layers of mudstone facies *F* (0.1–1.5 m thick) in the fourth, uppermost cycles.

The above discussed successions of alternating facies *SFm* and *F* indicate the presence of intense but short-term deposition of *SFm* facies sandstones in crevasse splay channels. When the high-energy conditions of *SFm* sandstones terminated and the flow ceased, mudstones were accumulated from suspension. They filled successive small crevasse channels.

In the Mielnik IG 1 section (Fig. 15B), the lowermost floodplain facies succession *F→SM→SC→SFm* reflects deposition in crevasse splay channels. Another suite of facies *SFr→SM→F→SFm* is a record of water overflow from the river channel onto the floodplain and initially deposition of ripples from bottom traction transport. The large thickness (2 m) of *SFr* facies sandstones suggests a long-term episode of ripple deposition. After a period of low-energy deposition, there were two episodes of high-energy deposition recorded in facies *SM* and *SFm*. They were associated with deposition in crevasses splay channels. The last mentioned suite *SFm→F→SFm→F* is composed of thin rhythms showing normal grain-size grading. Each rhythm consists of fine-grained sandstone facies *SFm* and massive mudstone facies *F*. These deposits were accumulated during flood periods in the channels of crevasse splays. Deposition of sandstone facies *SFm* corresponds to an initial, high-energy phase of crevasse splay formation. Mudstone facies *F* formed when the flow in the crevasse channels ceased.

The evolution of anastomosed system rivers is determined by overbank processes which include avulsion and deposition of a floodplain. Avulsion processes in the anastomosed system rely on a change in flow direction from an active river channel towards the floodplain due to crevassing. This process leads to the formation of a new channel along the lowest segment of the floodplain. Frequent avulsions in floodplains result in the formation of an anastomosed river system represented by active and abandoned river channels (e.g. Smith, 1983; Tornqvist, 1993).

Crevasse splays are a very characteristic depositional landform of floodplains in anastomosed system rivers (Fig. 7B). The formation of crevasse splays on the floodplain occurred as a result of autocyclic factors, mainly strong flood surges in the river.

Thicknesses of floodplain facies were controlled by the relationship between the chanel and its base level. Rises of base level associated with fluctuations of relative sea level can lead probably to long-term vertical accretion of floodplain deposits (Cotter, Driese, 1998). An example of this scenario is the large thickness of the crevasse splays deposits with from the Radzyń IG 1 and Święcica 1 sections.

GROUP OF ESTUARINE DEPOSITIONAL SYSTEMS

Bay-head delta depositional system

Description. The characteristic feature of the bay-head delta depositional system (Fig. 8C, D) is the presence of very coarse-grained facies *Gm*, *Gl*, *SCp*, *SC*, *SCl*, *SCm* and *SMl*. The deposits form 2 to 9 m thick beds. The coarsest-grained facies were encountered in the Radzyń IG 1 (Fig. 17) and Podedwórze IG 2 (Fig. 18) sections. Fine-grained deposits, represented mainly by sandstones of facies *SFh*, *SFf* and *SFm*, and mudstones of facies *F* occur in the middle part of the deltaic succession in the Parczew IG 10 (Fig. 19) and Kaplonosy IG 1 sections (Fig. 20). The bay-head delta facies assemblage in these four sections, overlies sediments of various zones of the tidal flat within the estuarine system, and underlies nearshore deposits of the open-coast depositional system.

Interpretation. Packages of coarsening-upward deposits from the transitional fluvial-tidal succession of the Radzyń IG 1, Parczew IG 10, Podedwórze IG 2 and Kaplonosy IG 1 sections form a prominent complex of coarse-grained deposits overlying the estuarine tidal sequence and underlying the fine-grained series of open-coast (Figs. 17–20). The coarsening-upward facies sequences are commonly composed of distinct cycles. Three cycles are observed in the Radzyń IG 1, two cycles in the Podedwórze IG 1 and Parczew IG 1, and one cycle in the Kaplonosy IG 1 section (Fig. 21B). Coarse quartz grains within the succession Podedwórze IG 2 are very mature (Fig. 21A). This feature suggests relatively

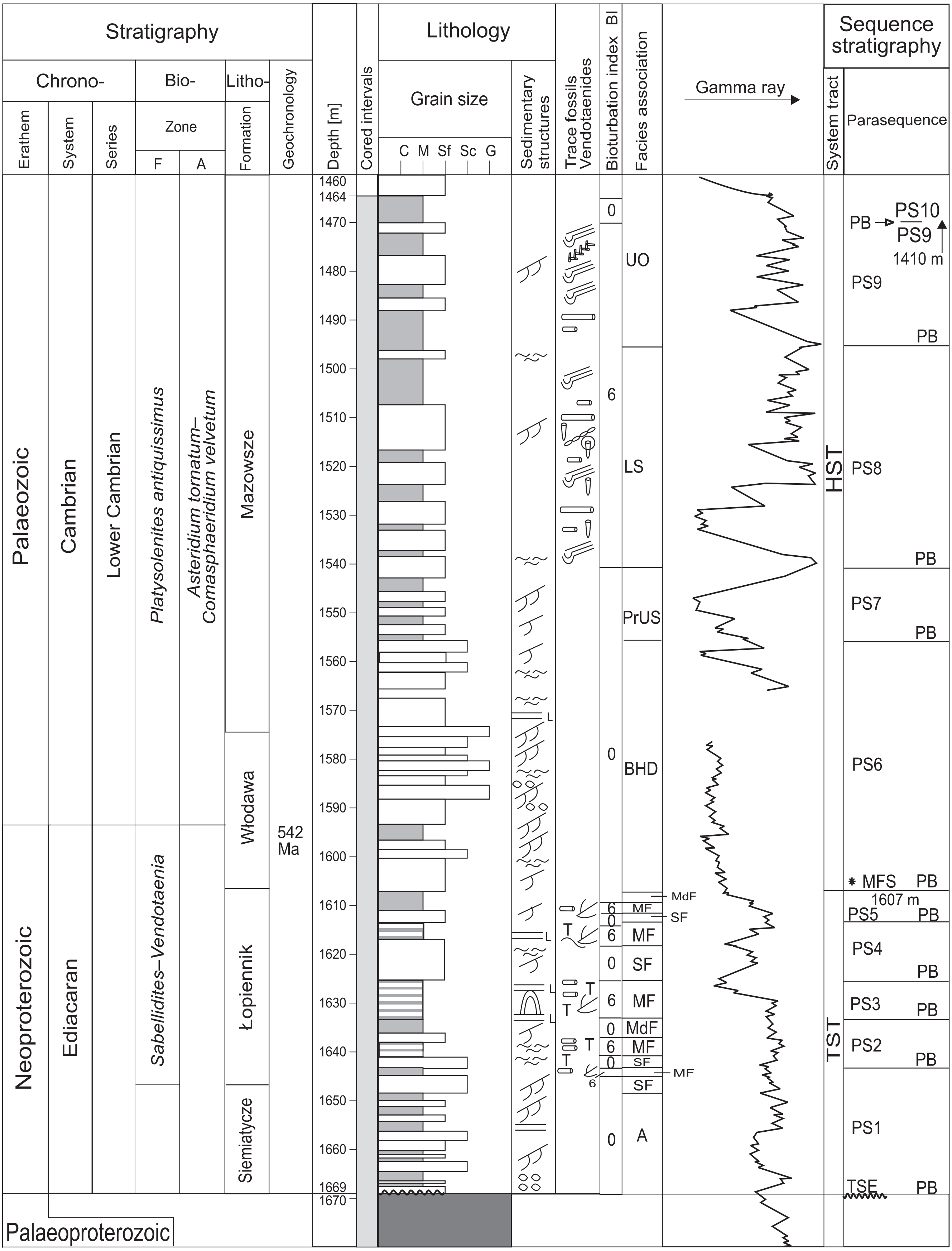

Fig. 17. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Radzyń IG 1 section

Explanations for figures 17–20, 22–25, 39, 41, 42

LITHOLOGY

Grain size

C claystones

M mudstones

Sf fine-grained sandstones

Sm middle-grained sandstones

Sc coarse-grained sandstones

G conglomerates

sandstone–mudstone–claystone heteroliths

volcanigenic rocks

crystalline basement

lithology non interpreted – non cored interval and lack of geophysical log

Sedimentary structures

bimodal cross-bedding

planar cross-bedding or indeterminate

ripple cross-lamination

horizontal stratification

L horizontal lamination

flaser lamination

lenticulare lamination

wave lamination

reactivation surface

mud drapes

mudstone and claystone clasts

erosional surface at the bottom of sandstone bed

erosional surface

BHD diastem bay-head delta diastem

FOSSILS

Vendotaenia antiqua

BIOTURBATION INDEX

0	no bioturbation recorded
1	sporadic
2	sparse
3	numerous
4	very numerous
5	intensive
6	totally bioturbation of deposits

TRACE FOSSILS

Monocraterion isp.

Bergaueria major

Bergaueria isp.

Skolithos linearis

Treptichnus triplex

Treptichnus bifurcus

Treptichnus isp.

Neonereites uniserialis

Gyrolithes polonicus

Telchichnus rectus

Planolites beverleyensis

Planolites montanus

Palaeopascichnus delicatus

T *Torrowangea rosei*

Gordia isp.

FACIES ASSOCIATIONS

BS	sand barrier
CB	central bay
BHD	bay-head delta
BR	sand bar
CST	subtidal channel
CACS	complex of amalgamated subtidal channels
CTF	tidal flat channel
IMS	inter-channel mud shoal
SF	sand tidal flat
MF	mixed tidal flat
MdF	mud tidal flat
PrUS	prograding upper shoreface
LS	lower shoreface
UO	upper offshore
A	alluvial deposit systems

Explanations for figures 17–20, 22–25, 39, 41, 42 – continuation

SEQUENCE STRATIGRAPHY

* MFS	maximum flooding surface
TSE	erosional transgressive surface
PB	parasequence boundary
TRS	tidal ravinement surface
SRS	shoreface ravinement surface
PB 7 *	parasequence boundaries on the basis of gamma ray log
PB → PS10 / PS 9 / 2052 m	parasequence boundary in the higher part of section
TST	transgressive systems tract
HST	highstand systems tract
LST	lowstand systems tract

STRATIGRAPHY

Chronostratigraphy	Moczydłowska, 1991; Pacześna, 1996, 2008
Biostratigraphy	Lendzion, 1983 a, b (F – faunal zones); Moczydłowska, 1991(A – Acritarcha zones); Pacześna, 1996, 2008
Lithostratigraphy	Areń, 1982; Pacześna, 2008, 2010 (in press)
?	interval of probable, undefined Ediacaran-Cambrian boundary

long transport, probably in a fluvial environment of high-energy braided rivers. The presence of pronounced coarsening-upward cycles repeated in a specific interval of the succession may suggest the deltaic depositional environment. The characteristic position of the deltaic succession at the top of estuarine deposits and beneath open coast series is most likely indicative of the presence of bay-head delta within this succession.

Central bay and estuary mouth depositional system

Estuarine central bay facies association

Description. Central bay association consists of facies *F*. Facies *F* are represented by dark grey and, rarely, black fine-grained mudstones and claystones of massive structure. They contain neither ichnofossils nor any other organic material, except for numerous Vendotaenides thallus. Accessory lithologic components are represented by irregular pyrite concretions. These facies are restricted to the southeastern part of the basin (Łopiennik IG 1, Białopole IG 1, Horodło 1 and Terebiń IG 5 sections) (Figs. 22–25). Deposits of these facies types form thick (2–14 m) packages there. In a vertical succession, the central bay association deposits occur in the lower parts of the estuary sequence, where they are interbedded with associations representing deposits of tidal flat (Fig. 26). They overlie the volcanic series (Figs. 22–25).

Interpretation. Central bay facies association is represented by the finest-grained sediments deposited through a rain of very fine-grained suspension from stagnant waters. This is a typical very low-energy association. Facies *F* contain no ichnofossils, whereby indicating a completely unfavourable environment for the existence of deposit-feeders typical of such environments. The producers of feeding-dwelling burrows might have been eliminated due to either deficiency of nutrients in bottom sediments or low level of oxygenation. However, mass occurrence of *Vendotaenia antiqua* forma *quarta* thallus in the black claystones and mudstones (Fig. 27A, B) probably excludes the possibility of nutrient deficiency. After remains of the organisms had been incorporated into the sediment and buried, they could be a sufficient source of food for the sediment-inhabiting deposit-feeders which would have left a record of their living activity. Considering this argument, it is probable that the lack of trace fossils in these deposits was likely due to anoxia of the sedimentary environment. The presence of pyrite supports this interpretation. Anoxic conditions developed in a partly protected embayment during initial transgression phases and marine inundation of an alluvial plain.

Subtidal sand bars facies association

Description. The facies association of subtidal sand bars consists of facies *SFh*, *SFp* and *SFr*. These deposits are represented by few 4–9 m thick beds observed in the lower part of the estuarine tidal succession. The occurrence of sand bar facies association in this succession is limited to very few intervals in the Łopiennik IG 1 and Białopole IG 1 sections located in the southeastern part of the basin.

Interpretation. Subtidal sand bars are elongated sand ridges parallel to the direction of tidal currents and they are often part of the so-called marine sand body complex (Dalrymple *et al*., 1992; Kitazawa, 2007). Occurrence of subtidal sand bars is typical of the lower intertidal and subtidal zone in meso- and macrotidal estuaries (Dalrymple *et al*., 1992; Mallet *et al*., 2000) (Fig. 8A). The occurrence of diagnostic tidal indicators such as mud drapes and bimodal cross-bedding seem to confirm the interpretation of infrequent sand bodies as subtidal sand bars located in the estuary mouth area. Another argument confirming the origin of these deposits as tidal

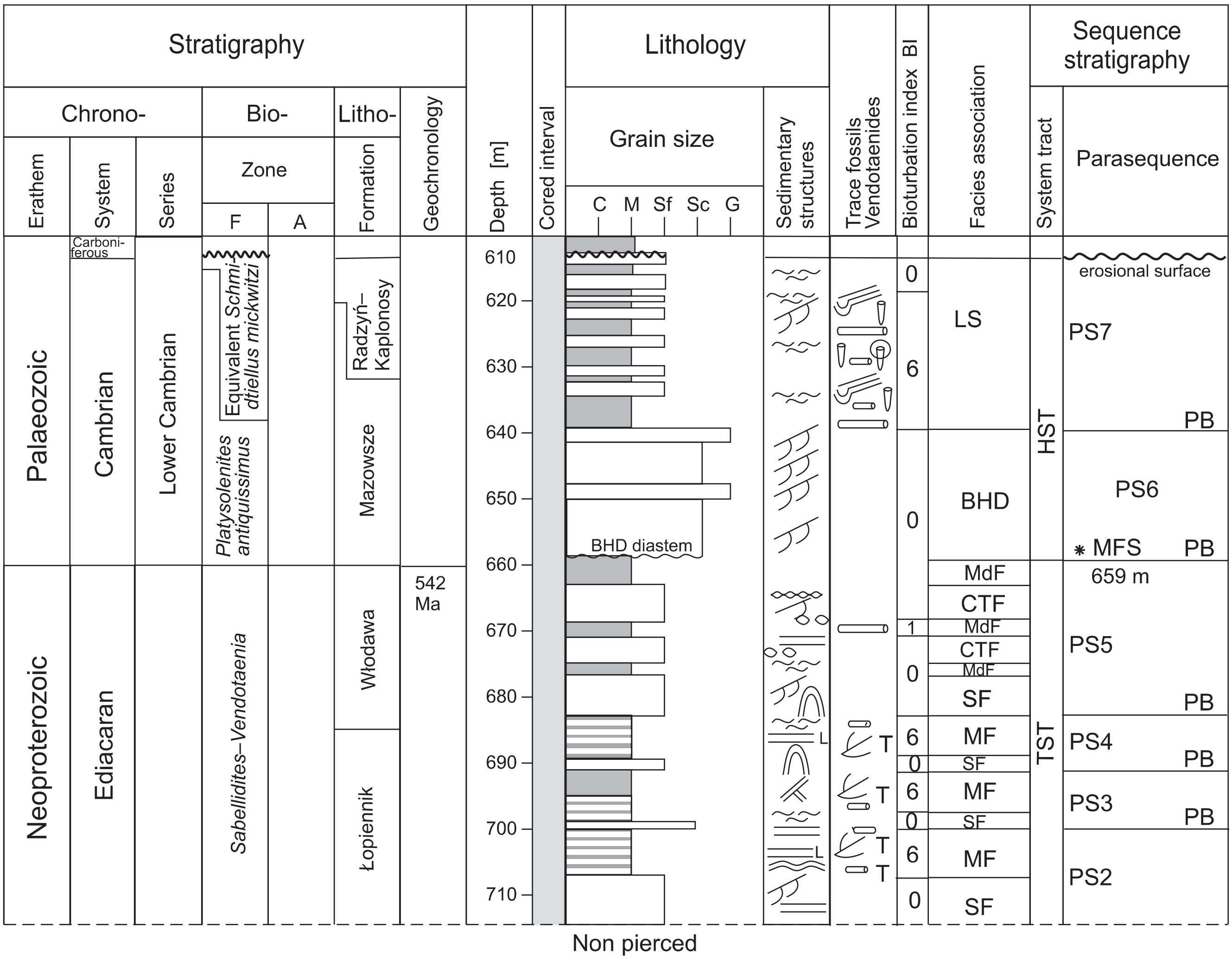

Fig. 18. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Podedwórze IG 2 section

For explanations see Fig. 17

sand bars is the position of the intervals within the estuarine succession under the complex of amalgamated subtidal channels and directly over the estuary's central bay deposits (Figs. 22, 25). Occasional sand bars, accompanied by subtidal channel deposits, probably also occur in the uppermost part of the estuarine complex. The limited occurrence of subtidal sand bars can be explained by destruction during the successive episodes of rapid relative sea-level rise during the transgressive stage of the Lublin estuary development.

Subtidal channels facies association

Description. This association consists primarily of facies *SCl*, *SFh*, *SFp*, *SFl*, *SCl* and *Gm*. Facies *SFm* and very fine-grained facies *F* occur infrequently. Trace fossils are sporadic and represented mainly by *Planolites montanus* Richter and *Planolites beverleyensis* (Billings) observed in thin interbeds of mudstone and claystone (facies *F*) or thickly laminated sandstone-mudstone heterolith (facies *FS*).

Interpretation. The analysed sections reveal intervals containing a distinctive vertical facies succession, very well manifested in the Łopiennik IG 1, Terebiń IG 5, Horodło 1, Białopole IG 1 sections (Figs. 22–25). Lower parts of most of the sequences contain very numerous mudstone and claystone clasts occurring above a well-developed erosional surface (Figs. 28, 29B, C). The clasts locally form thin, up to 0.3 m thick, intraformational muddy conglomerate layers of facies *Gm* (Fig. 29C). Above it, there is usually facies *SFl* represented by well-sorted fine-grained sandstones with low-angle (10–20°) planar cross-bedding, or, rarely, high-angle (25–40°) cross-bedded facies *SFp* or facies *SFr*. Such an alternation of low-angle and high-angle bedding with ripple cross-lamination is characteristic of tidal deposits (Abbott, 1998). There is also bimodal planar cross-bedding, supporting the hypothesis of tidal processes at that time (Fig. 29C).

Another evidence for tides is provided by the presence of mud drapes on the foreset of cross-bedding surfaces in sand-

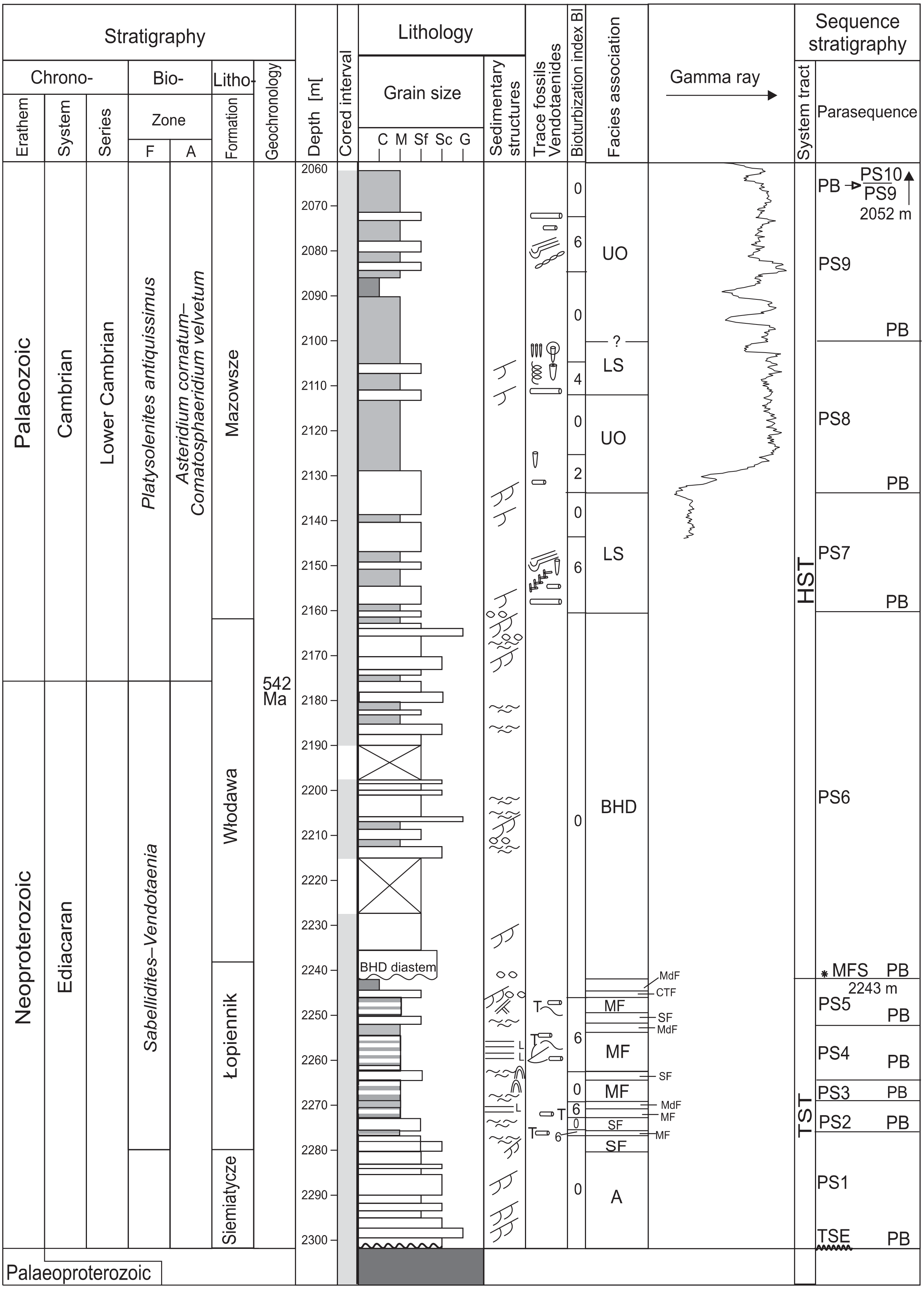

Fig. 19. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Parczew IG 10 section

For explanations see Fig. 17

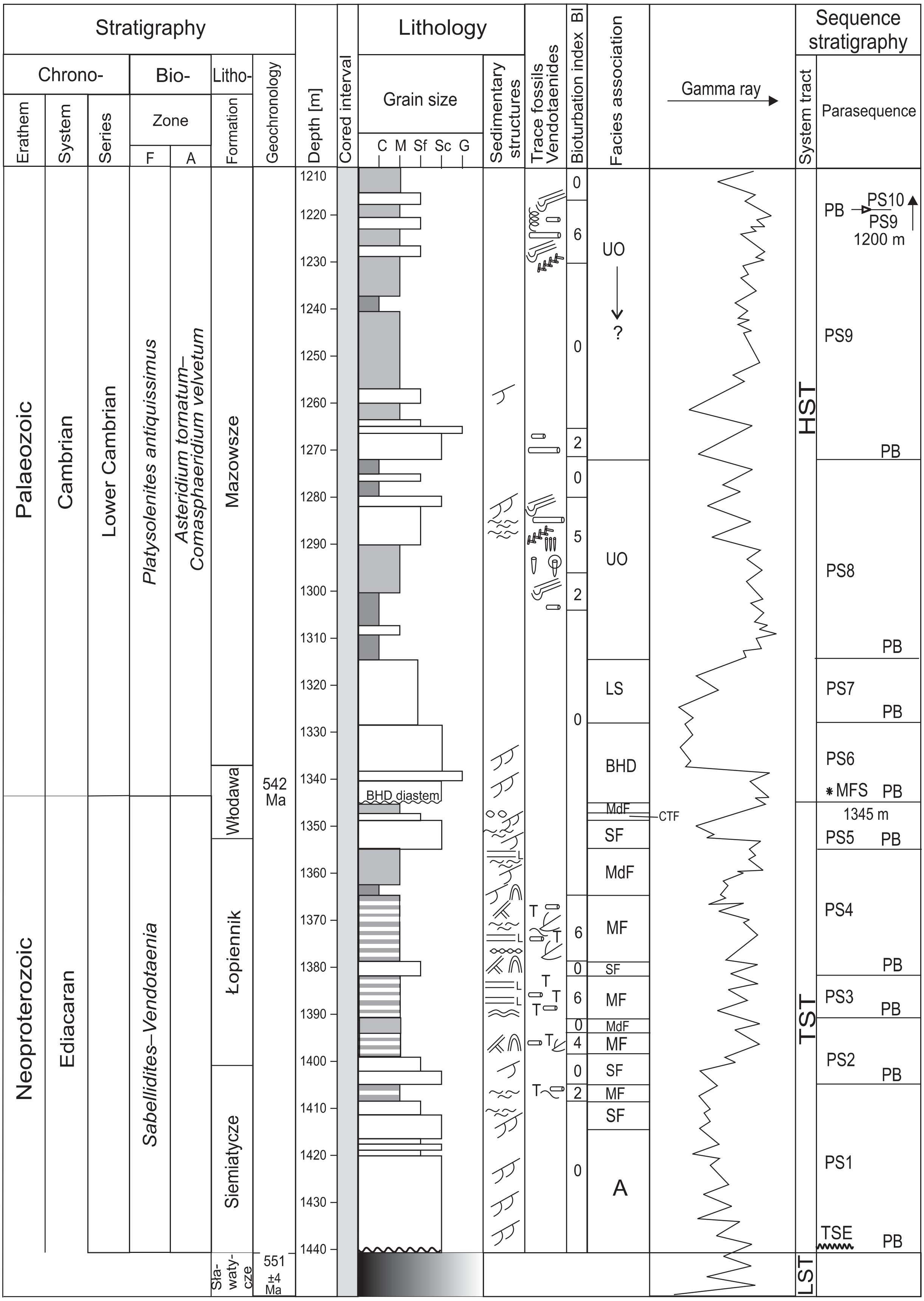

Fig. 20. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Kaplonosy IG 1 section

For explanations see Fig. 17

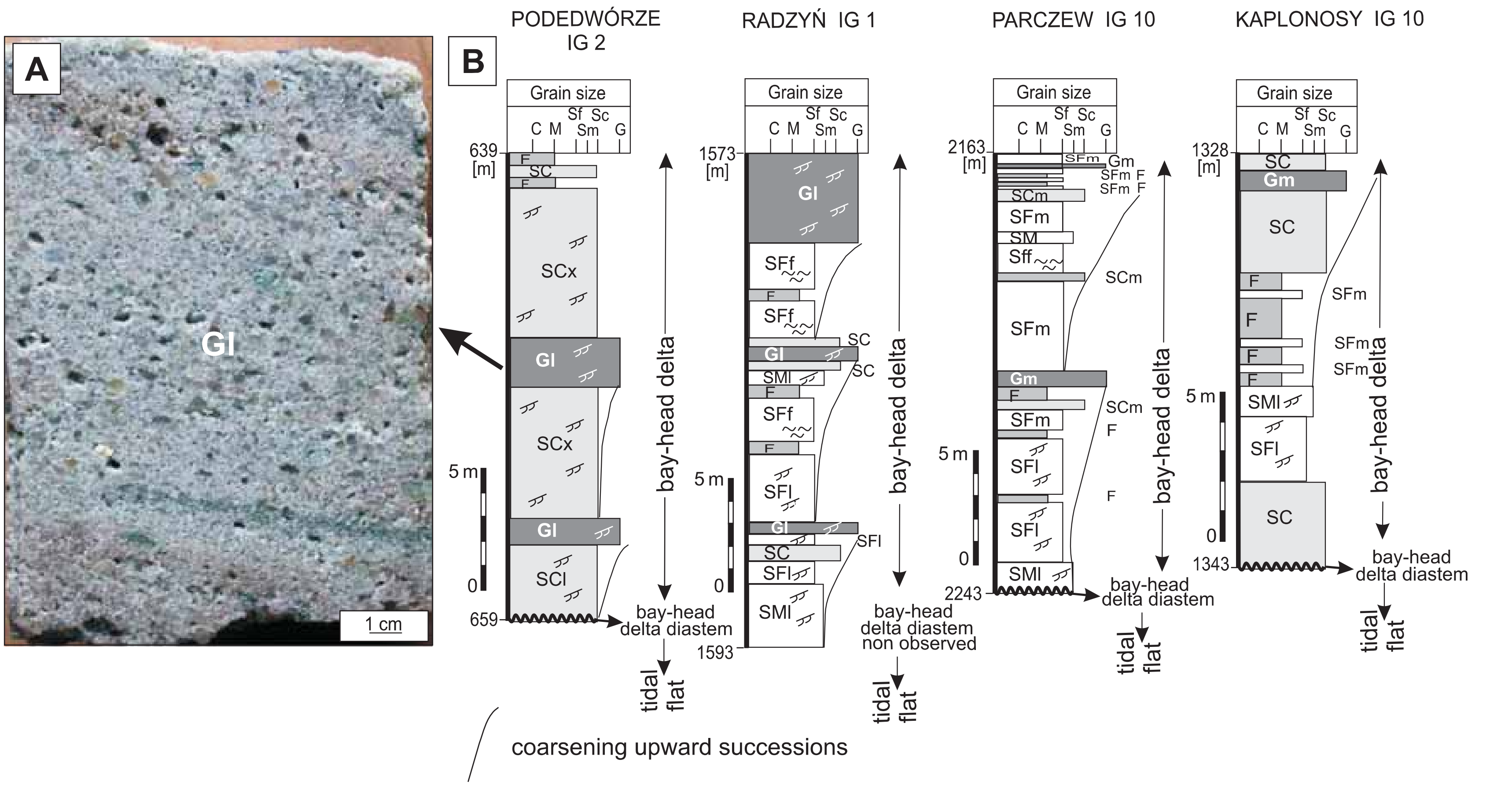

Fig. 21. Bay-head delta facies suites, early Lower Cambrian, Mazowsze Formation. A – monomictic quartz conglomerates and gravelstones with poorly visible low-angle cross-bedding, Podedwórze IG 2 borehole, depth 649.0 m; B – coarsening upward cycles, Lublin upper estuary

For explanations of grain size, sedimentary structures and facies code see Fig. 10

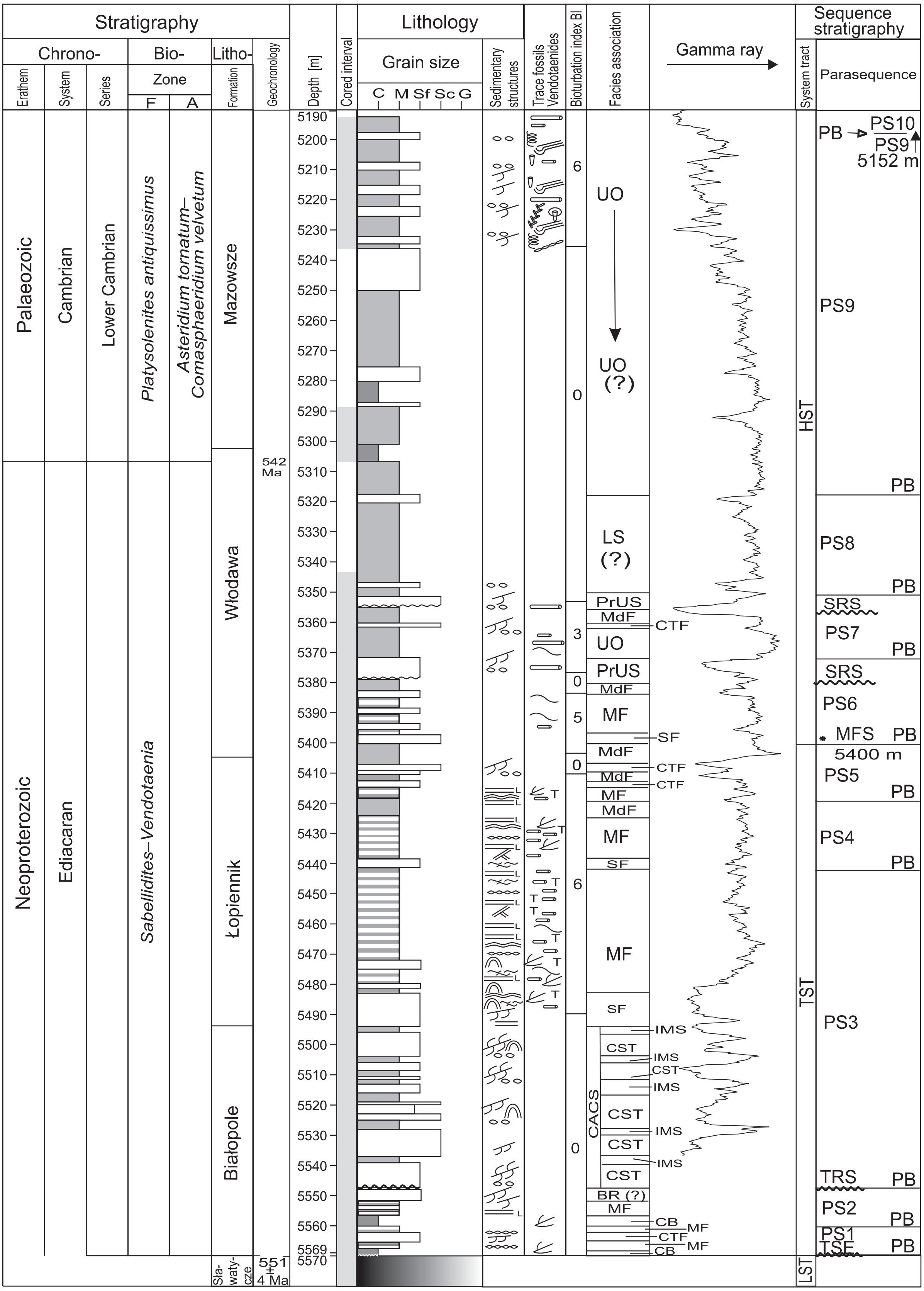

Fig. 22. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Łopiennik IG 1 section

For explanations see Fig. 17

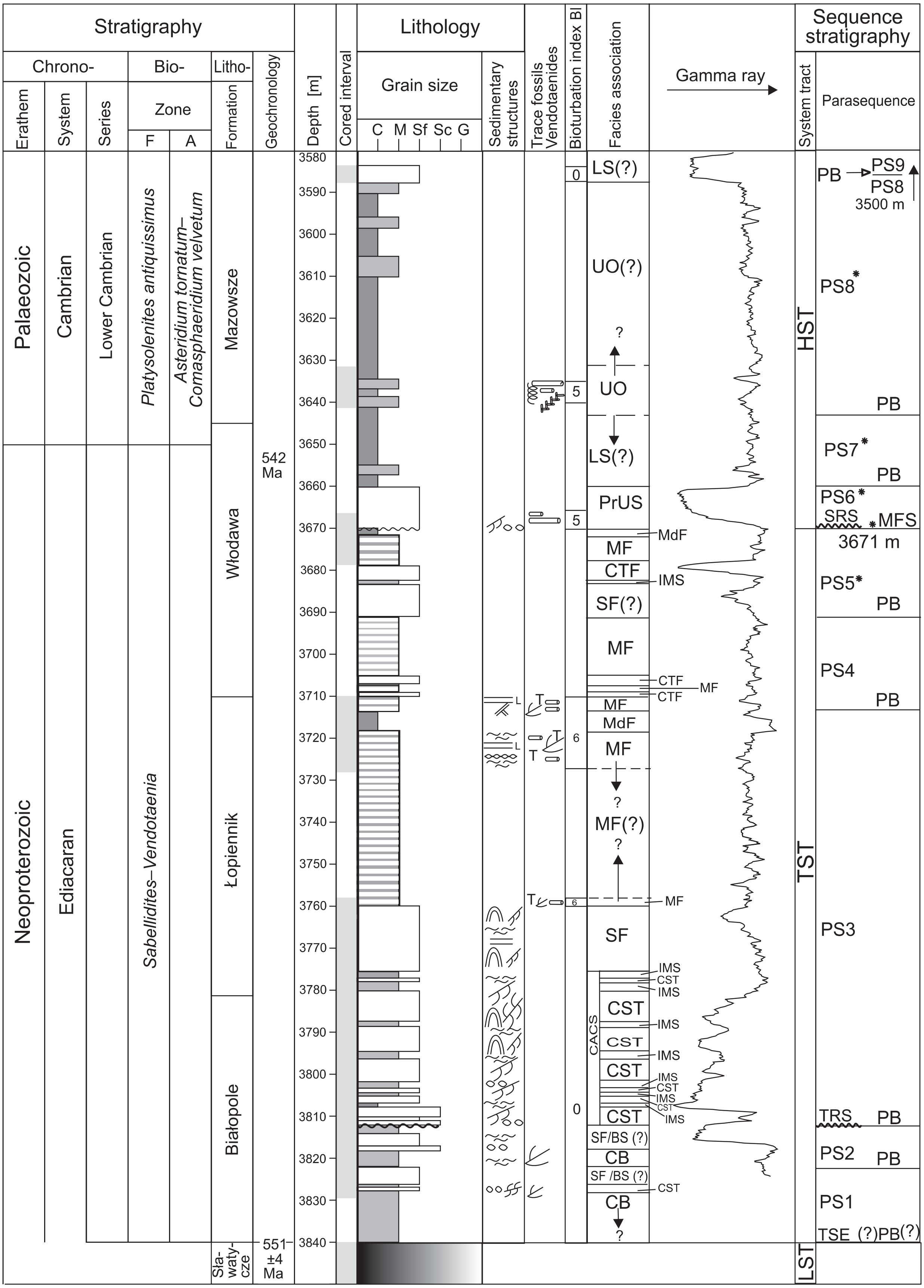

Fig. 23. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Terebiń IG 5 section

For explanations see Fig. 17

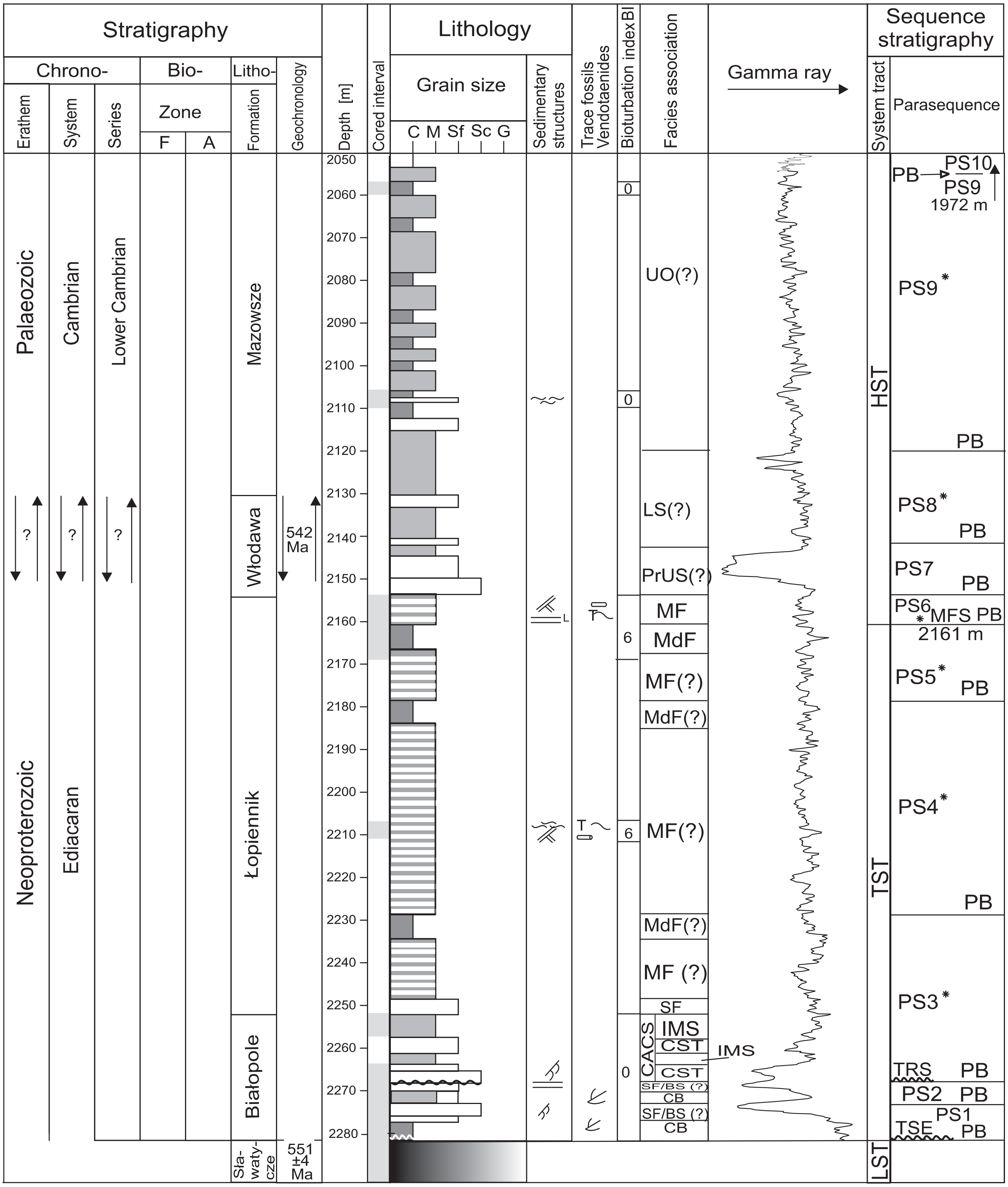

Fig. 24. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Horodło 1 section

For explanations see Fig. 17

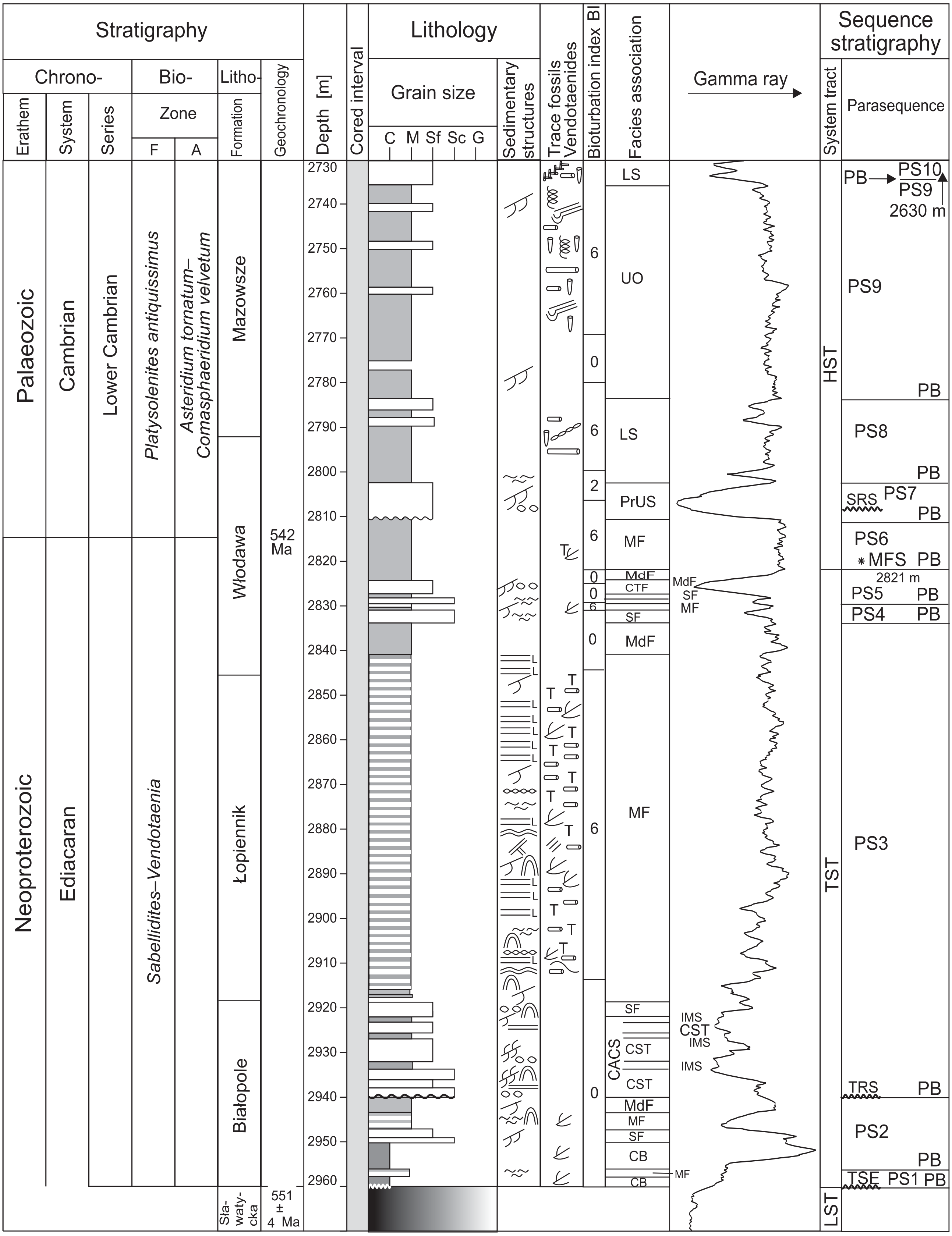

Fig. 25. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Lower Cambrian, Białopole IG 1 section

For explanations see Fig. 17

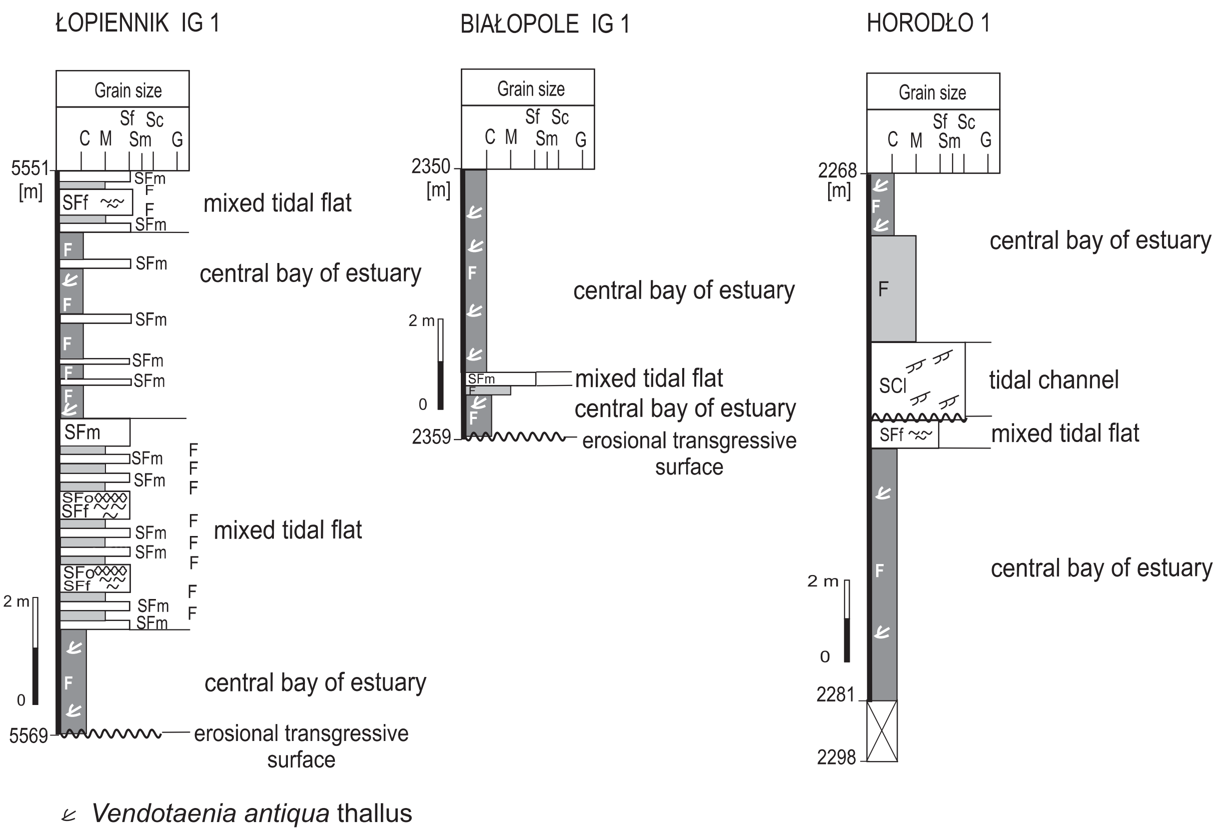

Fig. 26. Facies architecture of the central bay basin, lower Lublin estuary. Note presence of *Vendotaenia antiqua* thallus in black claystones

For explanations of grain size, sedimentary structures and facies code see Fig. 10

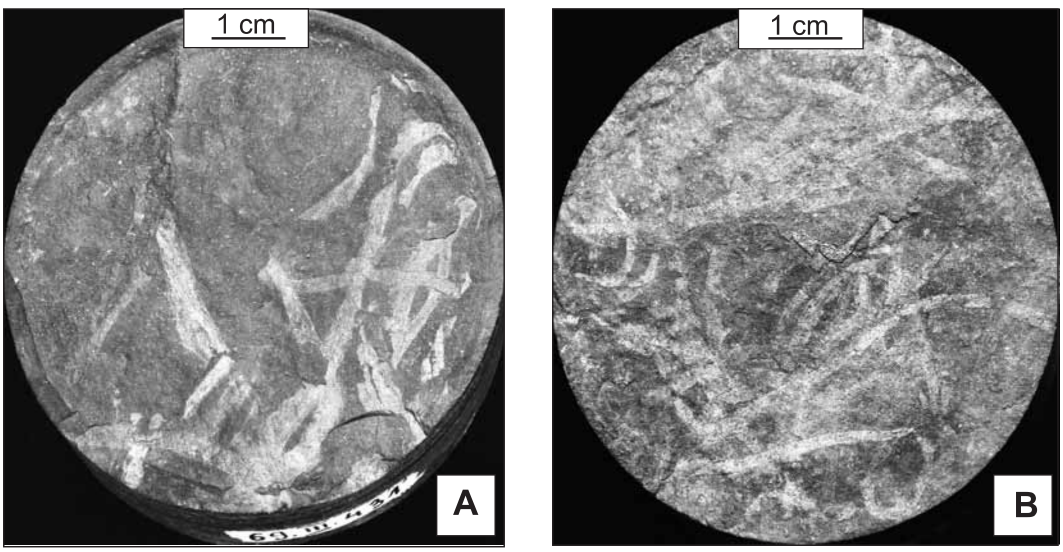

Fig. 27. Characteristic mass occurrence of Cyanobacteria thallus in black claystones of central bay facies association, late Ediacaran, Białopole Formation, Horodło Member, Łopiennik IG 1 borehole

A – *Vendotaenia antiqua* forma *quarta* Gnilovskaya, depth 5558.8 m; **B** – *Vendotaenia antiqua* forma *tertia* Gnilovskaya, depth 5551.6 m

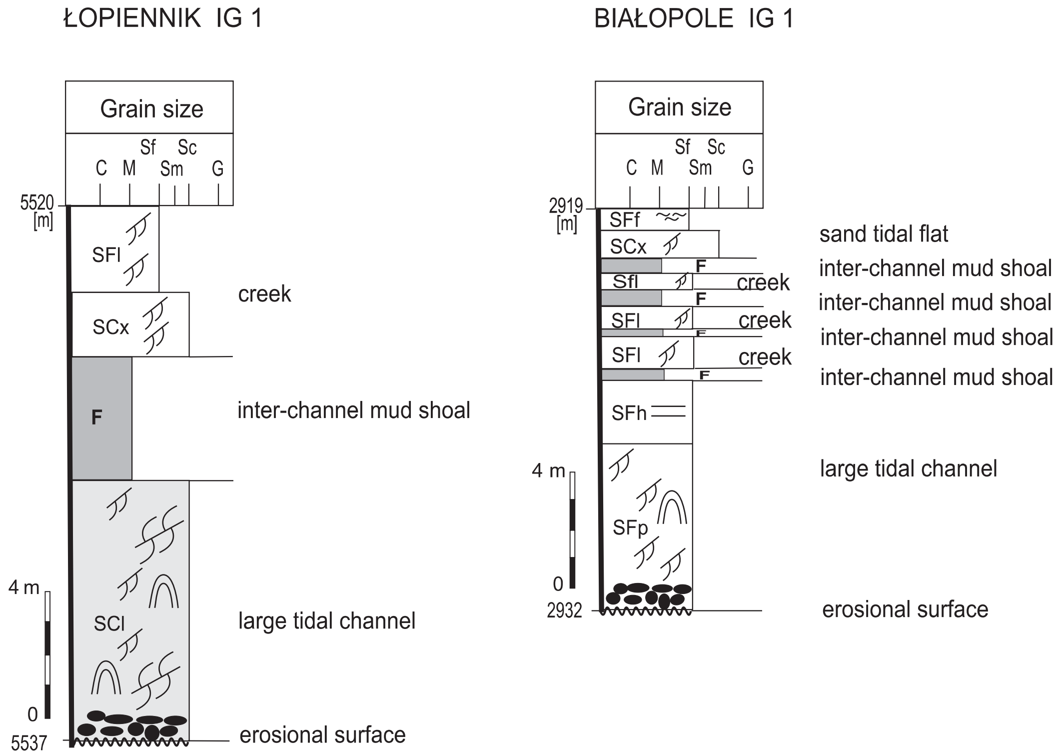

Fig. 28. Facies successions of variously sized subtidal channels and inter-channel mud shoals, Lublin lower estuary

For explanations of grain size, sedimentary structures and facies code see Fig. 10

stone facies *SFl* (Fig. 29B, C), especially well-developed in the Terebiń IG 5 section. The upper part of the packages is commonly represented by facies *SFl* grading into facies *SFr* or, rarely, facies *SFo*. Most of the packages show a fining-upward trend, with facies *F* occurring usually at their tops. The packages are 2 m to 14 m thick. In the Łopiennik IG 1, Horodło 1 and Białopole IG 1 sections, the subtidal channels facies association underlies the sandflat, mixed flat or mudflat facies association (Figs. 22, 24, 25). The subtidal channel facies association (and its position within the sections) indicates that it represents tidal channels developed in a permanently submerged subtidal zone (Fig. 8A).

The range of sedimentary structures suggests that the subtidal sediments were accumulated under high-energy conditions and dominance of mechanical reworking of the sediment. Such conditions probably resulted in a small content of organic matter within the deposit (Fillion, Pickerill, 1990). Scarce nutrient resources were the reason for low frequency of deposit-feeders represented by *Planolites montanus* Richter and *Planolites beverleyensis* (Billings) burrows observed in thin interbeds of mudstone (facies *F*). Sediment mobility prevented the filter-feeders from establishing burrows, and thus kept them away from the subtidal channel environment. Worth noting is also the lack of Vendotaenides thallus in mudstone and claystone packages of the subtidal channels association. It was likely due to unfavourable living conditions of highly variable environment of simultaneously migrating tidal channels and inter-channel mud shoals. The organisms were unable to inhabit such mobile environments.

Considerable variability in thicknesses of intervals corresponding to subtidal channels can indicate their different size. Thick sequences suggest the occurrence of main distributary channels, whereas thinner thicknesses are related to second order size channels (creeks) of the subtidal channels drainage network (Fig. 28).

An overlapping of successive thick sandstone and thin mudstone packages represented a large subtidal channels separated by inter-channel mud shoals, being a result of rapid, lateral migration of channels across the subtidal zone, especially across its lowest, sand part (e.g. Einsele, 2000). Each sandstones package in this complex represents large tidal channel which was active, before it was abandoned and subsequently covered by deposits of next channel. Sandstone packages in the complex are characterised by similar thicknesses of up to 9 m. A complex of amalgamated subtidal channels is particularly well developed in the south-east of the Lublin basin (Łopiennik IG 1, Terebiń IG 1 and Białopole IG 1 sections) (Figs. 22, 23, 25).

Inter-channel mud shoal facies association

Description. The inter-channel mud shoal association is represented only by very fine-grained mudstone facies *F*. The mudstones form 0.5 to 6.0 m thick beds.

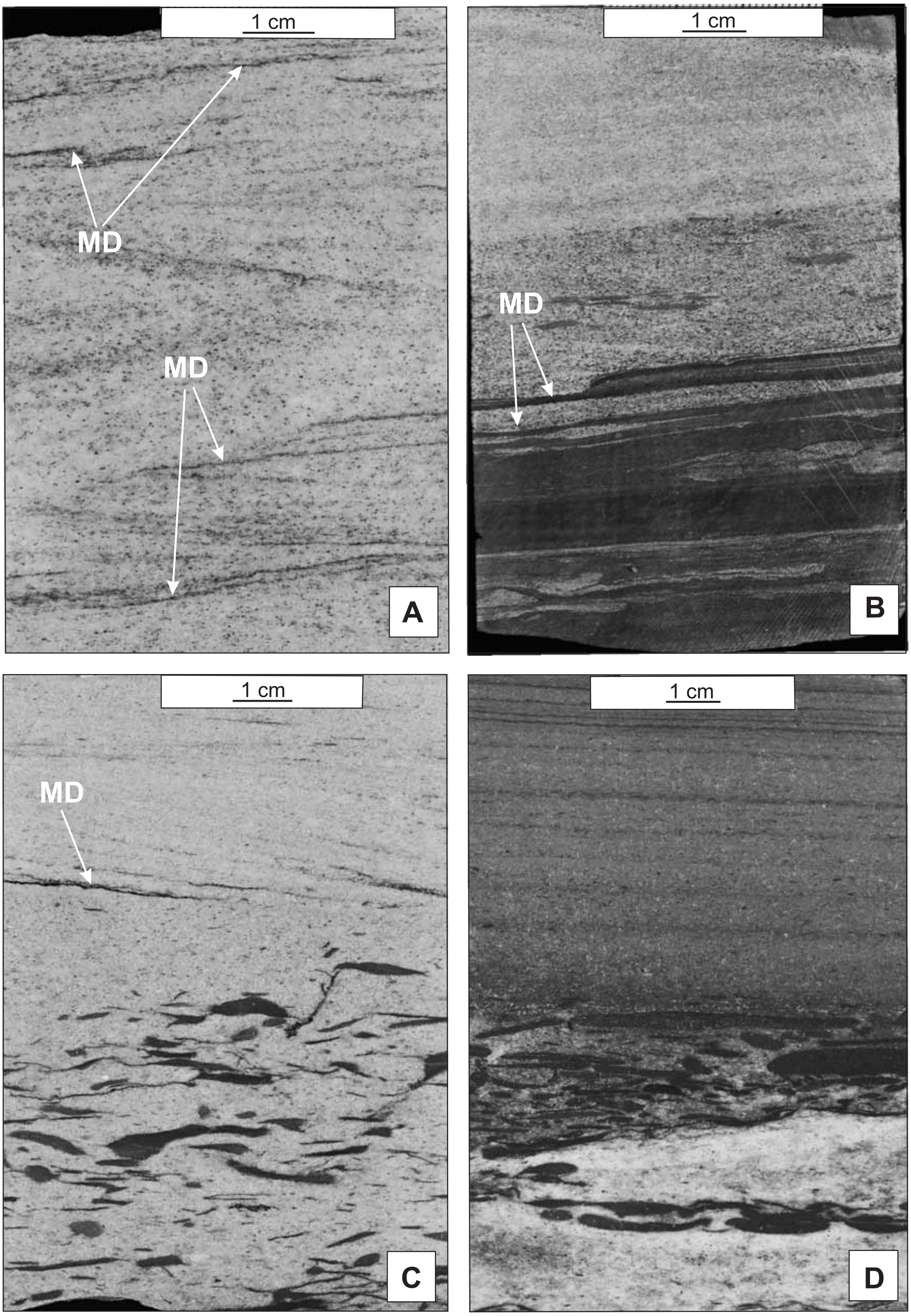

Fig. 29. Tidal channel deposits, late Ediacaran

A – bimodal low-angle cross-bedded fine-grained sandstone, MD – very thin mud drapes on the cross-bedding foreset. Tidal channel facies association on tidal flat. Busówno IG 1 borehole, depth 3655.0 m, Łopiennik Formation; **B** – low-angle cross-bedded fine-grained sandstone with mudstone intraclasts and thick mud drapes (MD). Subtidal channel facies association. Łopiennik IG 1 borehole, depth 5406.0 m, Łopiennik Formation; **C** – low-angle bedded fine-grained sandstone with claystone intraclasts. MD – thin mud drapes on cross-bedding foreset. Subtidal channel facies association. Białopole IG 1 borehole, depth 2922.5 m, Białopole Formation; **D** – at the base of core sample: large-scale low-angle cross-bedded sandstones with mudstone clasts overlain by horizontally bedded fine-grained sandstones. Tidal flat channel facies association. Łopiennik IG 1 borehole, depth 5379.0 m, Włodawa Formation

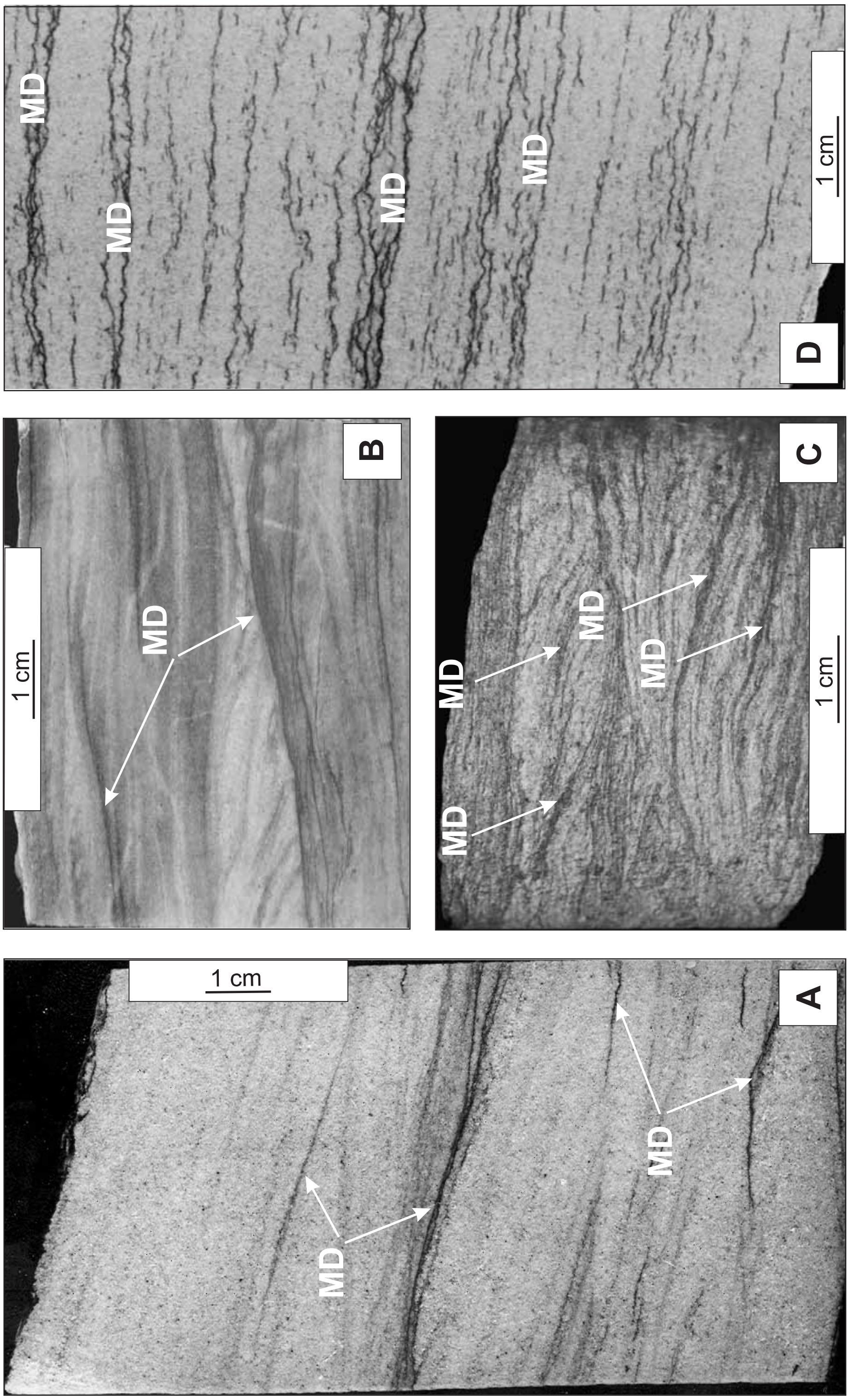

Fig. 30. Sand tidal flat deposits, sand tidal flat facies association, late Ediacaran

A – bimodal low-angle cross-bedded fine-grained sandstone. Note thin mud drapes (MD) on cross-bedding foreset. Białopole IG 1 borehole, depth 2920.0 m, Białopole Formation; **B** – ripple cross-lamination in fine-grained sandstones with thin mud drapes (MD) on cross-lamina foreset. Łopiennik IG 1 borehole, depth 5491.0 m, Łopiennik Formation; **C** – ripple cross-lamination in fine-grained sandstones. Note numerous thin mud drapes (MD) at the cross-lamina surface. Białopole IG 1 borehole, depth 2913.2 m, Łopiennik Formation; **D** – thin mud drapes forming subtle flaser lamination in very low-angle cross-bedded fine-grained sandstone. Białopole IG 1 borehole, depth 2922.2 m, Białopole Formation

Interpretation. Massive mudstones and occasional claystones of the inter-channel mud shoal association were deposited on the shoals extending between tidal channels (Yoshida *et al*., 2001). Their origin can be related to the overflow of water from channels. Mud and clay deposition occurred in a much lower-energy shoal environment, as compared to the deposition in tidal channels. Its occurrence is limited to the subtidal part of the tidal succession in the Łopiennik IG 1, Terebiń IG 5, Horodło 1 and Białopole IG 1 sections, located in the southeastern region of the basin (Figs. 22–25). An additional fact which facilitates distinguishing between the inter-channel mud shoal and mud flat deposits is the co-occurrence of the former with deposits of large tidal channels in the complex of amalgamated subtidal channels.

Tidal flat depositional system

Sandflat facies association

Description. The sandflat deposits consist of well-sorted fine-grained, rarely medium-grained, sandstones of facies *SFh*, *SFf*, *SFr* and *SF* with very rare thin mudstone and claystone layers occurring most frequently in the upper parts of the intervals. The sandflat deposits reach 2 to 16 m in thickness, with the average ranging from 4 to 8 m. The threshold value for distinguishing the sandflat in the intertidal zone (Fig. 8A) is the sand fraction percentage of >75% (after classification of Shepard, 1954 and Kim *et al.*, 1999), or 95% of the sand fraction assumed as the lower boundary, as observed on modern North Sea tidal flats (Hertweck, 1994).

Interpretation. The intertidal zone (Fig. 8A) consists of a specific repeated tripartite lithological suite of always the same packages of sandstones, horizontally finely laminated heteroliths and mudstones or claystones. It reflects decreasing flow energy of the environment and a change in the type of transport to suspension settling. These processes were responsible for the development of three facies associations within the intertidal setting. Sandflat facies association occupies the lowest part of the intertidal zone. Among sedimentary structures, the most frequent is ripple cross-lamination (Fig. 30B, C) and flaser lamination (Fig. 30D) proving deposition under lower flow regime (Zieliński, 1998). They are associated with the high velocity of tidal currents (Fillion, Pickerill, 1990). Cross-bedding is represented by a rare planar high-angle (30–45°) variety indicating deposition from saltation transport under lower flow regime. Bimodal cross-bedded sets are relatively common, which are indicators of oscillatory tidal currents on the sandflat (e.g. Nishikawa, Ito, 2000; Yoshida *et al.*, 2001) (Fig. 30A). The upper parts of the intervals containing sandflat deposits show increased amount of mud-clay sediments and occurrence of thin interbeds of sand-mud-clay heteroliths. It suggests a considerable decrease in energy. During low-energy periods, deposit-feeder burrows of *Planolites montanus* Richter are observed in thin claystone and mudstone layers, very rare in transition areas into the mixed flat. Events of increasing energy are indicated by the occurrence of horizontal stratification in fine-grained sandstones. Sandflat most commonly overlies either the complex of amalgamated subtidal channels or, more rarely, central bay and subtidal mud shoal (Figs. 20–23).

Mixed tidal flat facies association

Description. Among facies types, the most common is horizontally laminated heteroliths *Fh*. Facies *SFr*, *SFf*, *SFo*, *SFw*, *SFm* and *F* are rarer and occur as very thin layers within heterolithic facies *Fh* (Fig. 31). The trace fossil assemblage is abundant but taxonomically poorly diversified (Pacześna, 1996). According to Shepard (1954) and Kim *et al.* (1999), the content of sand fraction ranging from 25% to 75% is the basis for its discrimination. Another percentage limit of the sand fraction content is proposed by Hertweck (1994) for modern mixed flats – 55%. Mixed tidal flat association is the most widespread within the intertidal zone in the investigated basin.

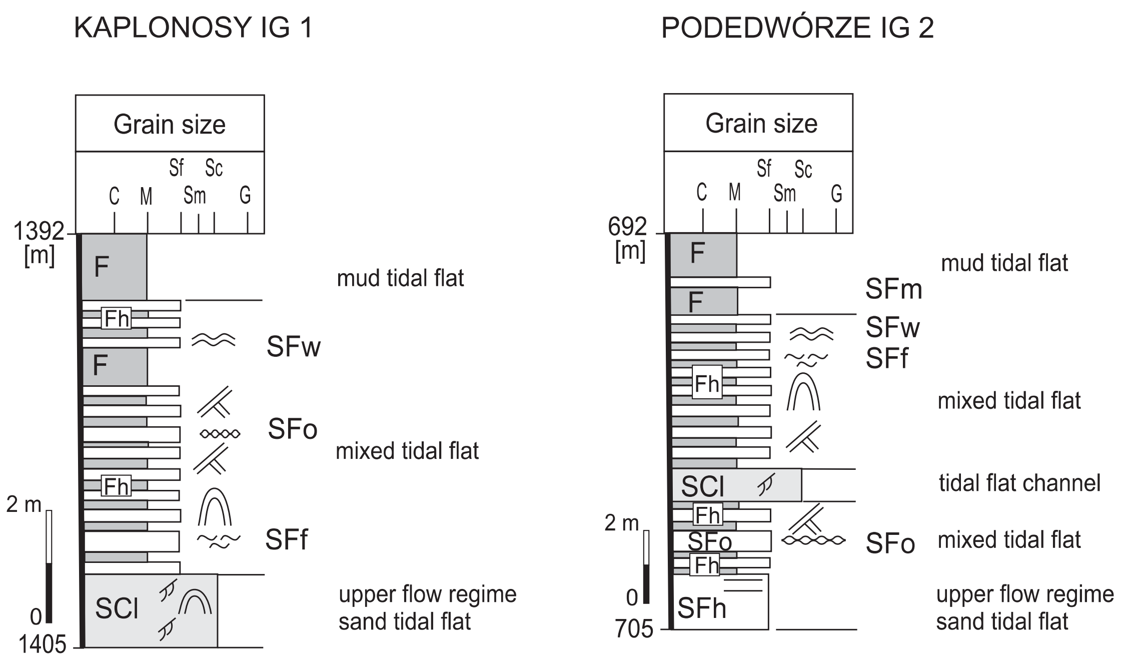

Fig. 31. Facies association of sand, mixed and mud flats, upper part of the Lublin estuary. Note characteristic tidal evidences such as mud drapes and reactivation surfaces in the finely laminated cyclic tidal rhythmites of facies *Fh*

For explanations of grain size, sedimentary structures and facies code see Fig. 10

Interpretation. Mixed tidal flat occupies an intermediate position between the sandflat and mudflat (Fig. 8A). The specific facies *Fh* is represented by alternating very thin (up to 1 mm in thickness) fine-grained sandstone lamina and mudstone, or rarely, claystone lamina (Fig. 32A, C). There are also thin (2–20 cm) interbeds of fine-grained sandstones with ripple cross-lamination, flaser, wave and lenticular lamination (Fig. 32E). Other facies types are thin fine-grained sandstones with lenticular lamination (Fig. 32E). The rarest sedimentary structure in the heterolithic facies is wavy lamination.

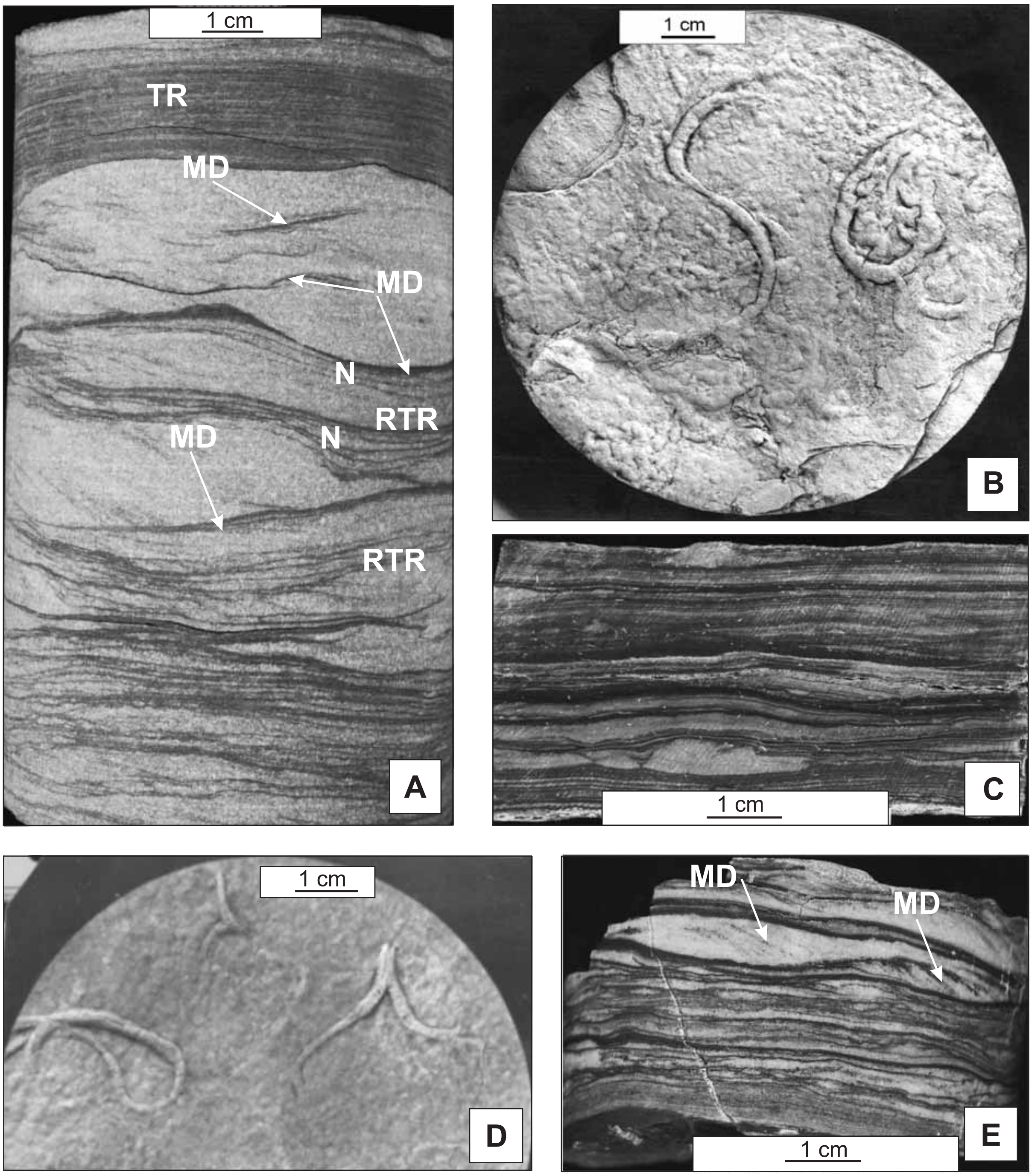

Fig. 32. Mixed tidal flat deposits. Facies association of mixed tidal flat, late Ediacaran, Łopiennik Formation

A – upper in the drill core sample: thin tidal rhythmite bed (TR) underlain by ripple cross-laminated fine-grained sandstone. Note numerous mud drapes on cross-lamina foreset (MD) and cyclic ripple tidal rhythmites with multiple mudstones laminae (RTR) representing a neap stage of neap-spring cycles. Białopole IG 1 borehole, depth 2919.2 m; **B** – *Torrowangea rosei* Webby, burrows of deposit-feeders in deposits of mixed tidal flat. Białopole IG 1 borehole, depth 2890.0 m; **C** – cyclic planar tidal rhythmites. Łopiennik IG 1 borehole, depth 5442.2 m; **D** – *Planolites montanus* Richter, burrows of deposit-feeders in deposits of mixed tidal flat. Łopiennik IG 1 borehole, depth 5451.2 m; **E** – tidal rhythmites with thin ripple cross-laminated fine-grained sandstone beds. Note very thin mud drapes (MD) on the cross-lamina surface representing spring stage of neap-spring cycles. Busówno IG 1 borehole, depth 3637.6 m

The presence of heterolithic facies suggests repeated episodes of different depositional processes. Sand material is transported and deposited by the tidal currents. Clay and mud material is deposited from suspension in stagnant or very slowly flowing waters. Heterolithic deposits of different proportions in thicknesses of sandstone and mudstone beds are considered tidal in origin, especially in those sections where other tidal features are present (e.g. Abbott, 1998; Kim *et al*., 1999; Yoshida *et al*., 2001). They also occur in other marine environments and on fluvial floodplains. Tidal origin of the sand-mud-clay heteroliths is supported by other tidal indicators observed in the underlying sandflat of the subtidal channel complex. These are: mud drapes (Fig. 33A, B), high frequency of reactivation surfaces (Fig. 33C, D) and bimodal cross-bedding. A reactivation surface is a curved surface that formed because changes in the tidal current formed a new pattern of strata or laminae, which differed from the previous pattern.

Thinly laminated heteroliths of facies *Fh* form a very distinctive package of tidal rhythmites (Fig. 32A, C, E). These are sand-mud-clay heteroliths, very regularly laminated. They form in modern settings, for example in the Bay of Mont-Saint Michel, Normandy (Larsonneur, 1994), and in the Fundy Bay, eastern Canada coast (Dalrymple, Zaitlin, 1994), under specific hydrological regime observed most often on macrotidal coasts in open funnel-shaped estuaries (Tessier, 1993). Ancient tidal rhythmites represented by thick packages have been reported from different geological systems (e.g. Alam, 1995; Singh, Singh, 1995; Porębski, 1995; Tessier *et al*., 1995; Gradziński, Doktor, 1996). They are frequently associated with macrotidal and seldom with mesotidal coasts. In the analysed sequences, tidal rhythmites form packages exceeding 100 m in thickness (Łopiennik IG 1) (Fig. 22). Tidal rhythmites contain a very peculiar set of trace fossils. Especially characteristic are deposit-feeder burrows represented by the *Planolites montanus* Richter (Fig. 32D) and *Torrowangea rosei* Webby (Fig. 32B) observed in facies *Fh*, with burrows forming close-packed clusters of constant burrow diameters below 1 mm. Abundance of trace fossils, that totally change the original structure of the sediment, indicates ecological opportunism of the tracemakers, constrained by severe environmental conditions, primarily by depletion of oxygen in the sediment and a wide range of salinity (Pacześna, 1996). Deposit-feeders inhabiting the mixed flat in feeding-dwelling burrows developed due to high tolerance to environmental stress. The large deposit-feeder population was also associated with abundance of nutrients, which is often related to low-oxygen environments. The characteristic features of burrows of opportunistic deposit-feeders that form densely packed clusters are currently considered one of the most sensitive indicators of brackish environments (e.g. Beynon, Pemberton, 1992; Pemberton *et al*., 1992a; Pemberton, Wightman, 1992a). The *Planolites montanus* Richter burrows are accompanied by very numerous representatives of the ichnogenus *Torrowangea rosei* Webby. Relatively frequent repichnia are represented by *Gordia* isp. suggesting sediment stability which allows epibenthic organisms to move on the surface. Numerous and diversified trace fossil assemblages is the typical feature of ancient mixed flats, commonly observed in tidal deposits of various geological systems. The Ordovician mixed tidal flats from Newfoundland (Fillion, Pickerill, 1990) and the Triassic tidal flats from British Columbia, Canada (Zonneveld *et al*., 2001) are good examples. The abundant but ichnotaxonomically poorly diversified trace fossil assemblage from the mixed flat of the Lublin–Podlasie basin can suggest unfavourable ecological conditions due to a low oxygen content in sediments and bottom water on protected estuarine tidal flats. Such environmental conditions forced the organisms to accept an opportunistic strategy of life, that resulted in their strong quantitative development and low ethological diversity. In the ichnological record, it is reflected by homogenization of mudstone and claystone structures entirely reworked by deposit-feeders, with closely packed *Planolites montanus* Richter and *Torrowangea rosei* Webby burrows 1 mm in diameter. Simultaneously, the great number of infaunal representatives was due to abundance of nutrients in the mud. Organic matter that originated from decay of large masses of Vendotaenides thallus on the tidal flat was probably the main component of the nutrients.

Mud tidal flat facies association

Description. The succession of mudflat is characterised by the finest grain size facies *F* (Fig. 34) and very rare, thin, fine-grained sandstone layers of facies *SFr*. The thickness of mudflat packeges ranges from 2 to 23 m.

Interpretation. The uppermost part of the intertidal succession (above the mixed flat) (Fig. 8A) is composed of the mudflat in all the sections. Mudflats are areas of clay and mud sedimentation related to high tides. The trace fossil assemblage is sparse and poorly ethologically and ichnotaxonomically diversified. During low tide, mudflat areas are subjected to subaerial exposure for a longer time than sand and mixed flat areas are (Hertweck, 1994). Variability of environmental conditions did not favour ethological diversity of infauna and epifauna. The trace fossils are represented by *Planolites montanus* Richter and *Gordia* isp. only.

Tidal flat channel facies association

Description. This association is represented mainly by facies *Gm*, *SCl*, *SC*, *SCx*, *Sfl*, *SFh* and *SFp*. Facies *SFr*, *SFf* and *F* are rare. In some sections the characteristic vertical facies succession starts with facies *Gm* composed of very thin, up to 0.10–0.20 m thick beds of intraformational muddy conglomerates. It is overlain by sandstone (facies *SFl*) (Fig. 29A), *SFp* or *SCl*, grading upwards into facies *SFh*. The most common channel fills include thin beds of a mudstone clasts conglomerate (facies *Gm*), low-angle cross-bedded coarse-grained sandstones (facies *SCl*) passing up to high-angle cross-bedded fine-grained sandstones (facies *SCx* and *SFp*) or horizontally stratified fine-grained sandstone (facies *SFh*). Upper parts of some sequences are represented by facies *SFf* (Fig. 35). These facies are arranged in fining and thinning upward packages, 0.6–4.5 m thick.

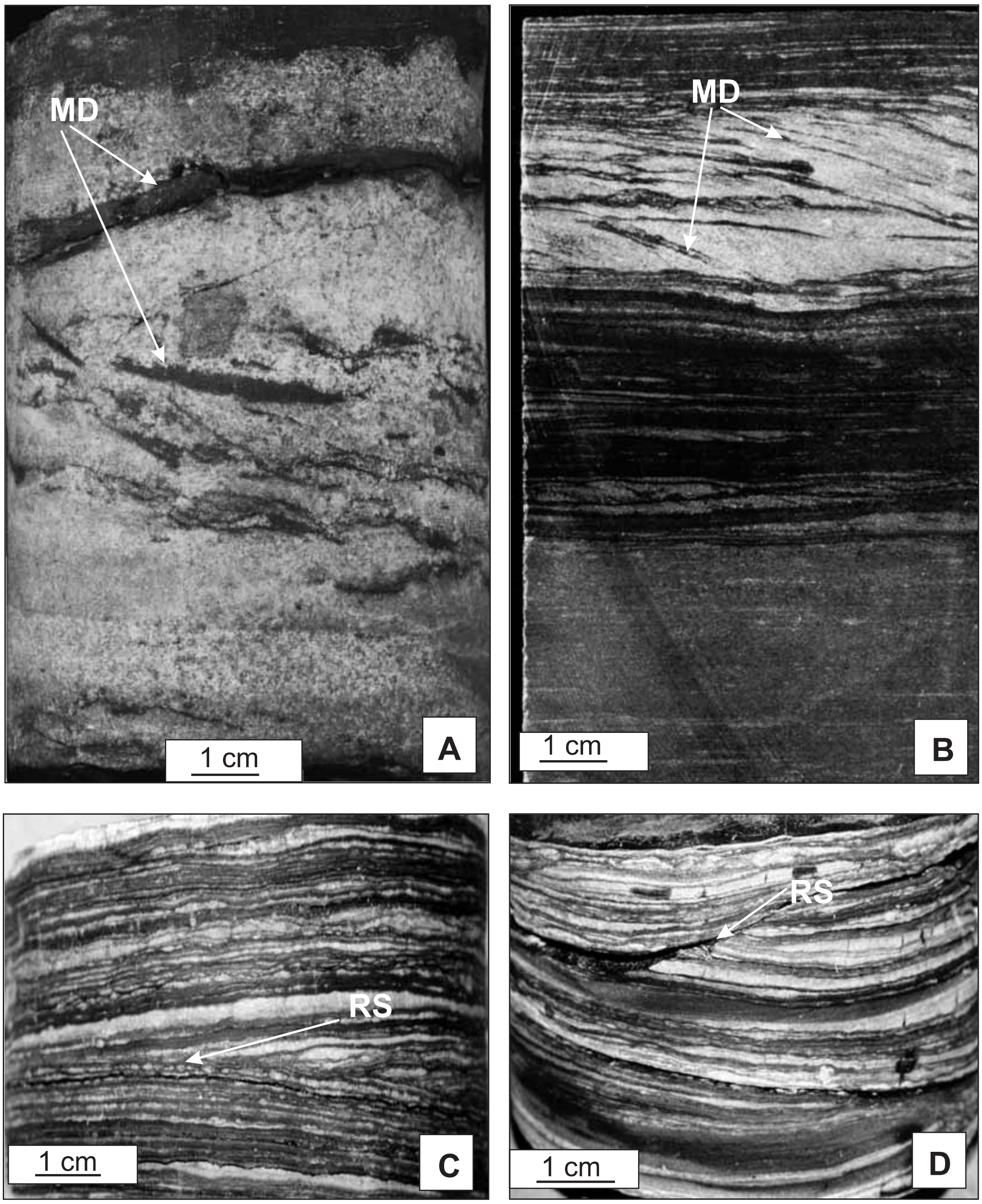

Fig. 33. Tidal evidences

A – thick mud drapes (MD) on the cross-bedding foreset in fine-grained sandstones. Tidal flat channel facies association. Terebiń IG 5 borehole, depth 3805.0 m, Białopole Formation; **B** – thin mud drapes on the cross-lamina surface in a thin fine-grained sandstone bed within a finely laminated sandstone–mudstone–claystone heterolith. Mixed tidal flat facies association. Łopiennik IG 1 borehole, depth 5474.4 m; **C** – reactivation surface (RS) in finely laminated rhythmites. Mixed tidal flat facies association. Łopiennik IG 1 borehole, depth 5467.4 m; **D** – reactivation surface (RS) in finely laminated cyclic rhythmites. Mixed tidal flat facies association. Podedwórze IG 2 borehole, depth 697.5 m

Interpretation. The tidal succession shows many features in common with a model succession of deposition in tidal channels that developed in the intertidal zone of the Ediacaran succession of the Lyell Land Group of the Eleonore Bay Supergroup from northeast Greenland (Tirsgaard, 1993). The model of sedimentation in tidal flat channels, created by Tirsgaard, involved the effect of environmental parameters different from the recent ones but specific of Ediacaran, pre-vegetational times. Tidal channels that developed on modern tidal flats commonly have a well-defined channels, meandering character, very well-developed fining-upward units and high contents of mud. Lack of vegetation and smaller mud content in the Ediacaran environments resulted in an increase of channel bank instability. These factors caused the formation of shallow and poorly defined channel morfologies of a braided character and lateral migration of channels across the tidal flat. Very fast vertical aggradation under high-energy flow took place in the Greenland channels

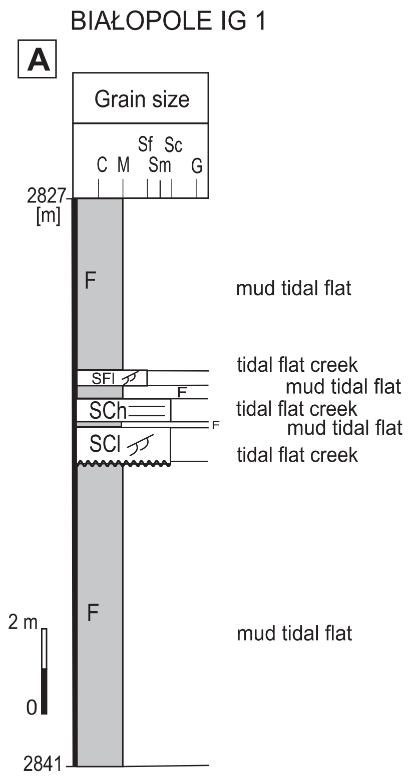

Fig. 34. Mud tidal flat facies association with tidal creeks in the lower (A) and upper estuary (B)

For explanations of grain size, sedimentary structures and facies code see Fig. 10

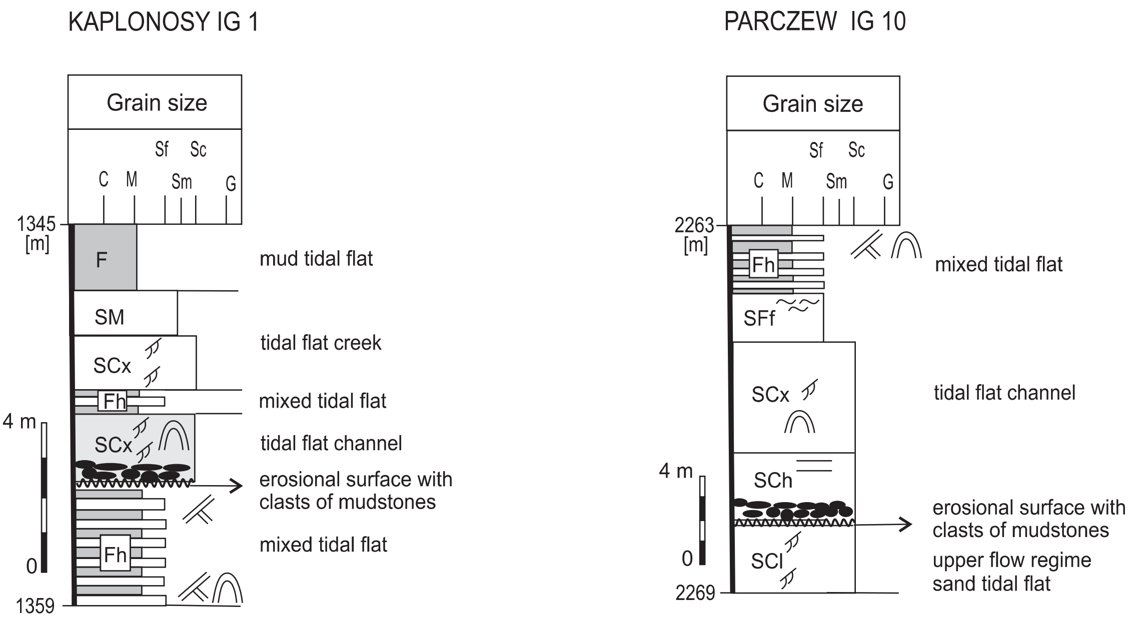

Fig. 35. Facies succession of tidal channels on sand and mixed flats, Lublin upper estuary. Note a presence of upper flow regime sand tidal flat in the Parczew IG 10 section

For explanations of grain size, sedimentary structures and facies code see Fig. 10

(Hamberg, 1991; Tirsgaard, 1993). The features in common for tidal channel suits in both studied and Greenland's succession are the following: similar sedimentary structures sequence indicating high-energy sedimentation in channels, clear predominance of coarse-grained material of the channel fills and interbedding of sandstone channel-fills with heterolithic deposits of the mixed flat. Among the features differentiating both these sequences is, first of all, a different structure of their channel fills. Greenland's channels are characterised by multistorey sand sheets. Individual sand sheets are separated by erosional surfaces. The studied channels have a less complicated structure and most of them are composed of a single sand sheet. It suggests weaker vertical aggradation of the sediment in the channel. In contrast to the Greenland's channels, the analysed deposits are interlayered not only with mixed flat but also with sand and mud flat sediments. It is especially worth noting that the channel deposits are relatively frequently interlayered with mud flat deposits. It indicates a greater content of mud material, which allowed the development of the mud flat within the succession.

Compared to subtidal channels, very low-relief erosional surfaces, small thickness of channel fills and the presence of thin basal lag beds suggest (Fig. 29D) relatively small-scale and shallower tidal channels located in a more proximal position within an estuarine tidal flat complex. The exclusive presence of channel packages in the upper portion of the tidal succession, corresponding to the intertidal zone, especially distinctly interlayered with sand, mixed and mud tidal flat deposits (Figs. 22–25) implies deposition in tidal channels developed in the intertidal zone.

Frequent ripple cross-laminated intervals observed in the upper part of the tidal channel complex indicate deposition from rhythmical bed transport and suggest a fall of energy flow in channels of the upper parts of mixed tidal flat at the transition to the lower mudflat. This is the so-called death tidal channel zone that occurs at the mudflat/supratidal zone boundary (Borrego *et al*., 1995). Because the analysed sections show incomplete tidal flat sequences, most of the channels ended within the upper part of the mixed flat. None of the analysed sections contain supratidal deposits overlying facies packages of the tidal flat. The lack of supratidal facies may have been due to removal of deposits by erosion during the five successive transgressive events which flooded the tidal flat. Another explanation for the absence of the supratidal zone is that it did not develop on both open and protected Neoproterozoic tidal coasts (Deynoux *et al*., 1993). Its absence can also be explained by periods of increased subsidence rate in the basin, exceeding sedimentation rate on the tidal flat (e.g. Alam, 1995). Above the maximum flooding surface, the tidal complex is overlain by the prograding upper shoreface packages.

OPEN COAST DEPOSITIONAL SYSTEM

Upper shoreface facies association

Description. The upper shoreface facies association is featured by the presence of coarse- and fine-grained sandstone facies *SCl*, *SFh* and *Sx*. Facies *SFp* is subordinate. Mudstones are rare and represented by facies *F*. The trace fossils occur in very small amounts.

Interpretation. The upper shoreface is an open-marine coast area located in the high energy surf zone and lies landward of the breaker zone (MacEachern, Pemberton, 1992) (Fig. 9). This zone is dominated by wave driven currents paralleing to shoreline and currents generated by tranlatory flow associated with plunging wave. The upper shoreface is characterised in the study area by predominant sandstone intervals containing low-angle planar cross-bedding, interbedded with intervals containing high-angle planar cross-bedding. This is a record of migrating multi-directional sinusoidal magaripples (*op.cit*). Strong sediment mobility in the upper shoreface zone, due to the process of bedform migration, enabled inhabiting of sand deposits by filtrators dwelling in vertical or oblique burrows. Due to stress-inducing living conditions for the tracemakers, the ichnological spectrum of the upper shoreface facies association is very poor. Trace fossils are represented by filter-feeders' domichnia *Monocraterion* isp. It is due to environmental conditions unfavourable for the preservation of dwelling burrows. Such conditions included mainly high environmental energy and predomination of cohesionless properties of coarse-grained deposits of the prograding shoreface (Frey, Howard, 1988). On the other hand, low organic matter content, resulting in reduction of nutrients in such a high-energy environment, efficiently eliminated deposit-feeders from the habitat, dwelling within the sediment in horizontal feeding burrows. Distinctive feature is the presence of very well developed erosional surfaces at the base of fine-grained sandstone beds of facies *SFp* and *Sx*. Directly on the erosional surface or above in the sandstone, there are randomly distributed mudstone and claystone clasts. These packages represent the proximal upper shoreface prograding during the highstand.

Lower shoreface facies association

Description. The lower shoreface facies association includes facies *SFh*, *SFp*, *Sx* and *SFl*. The common sedimentary structure is ripple cross-lamination observed in sandstone facies *SFr*. Also frequent is flaser lamination represented in sandstone facies *SFf* and mudstone facies *F*. Facies *SFp* is the least common. Abundance and diversity of trace fossils are high.

Interpretation. The lower shoreface is situated between fair-weather (normal wave base) and maximum wave base (storm wave base) (Fig. 9). This is an area of dominant wave, high-energy conditions. Most sedimentary structures reflect

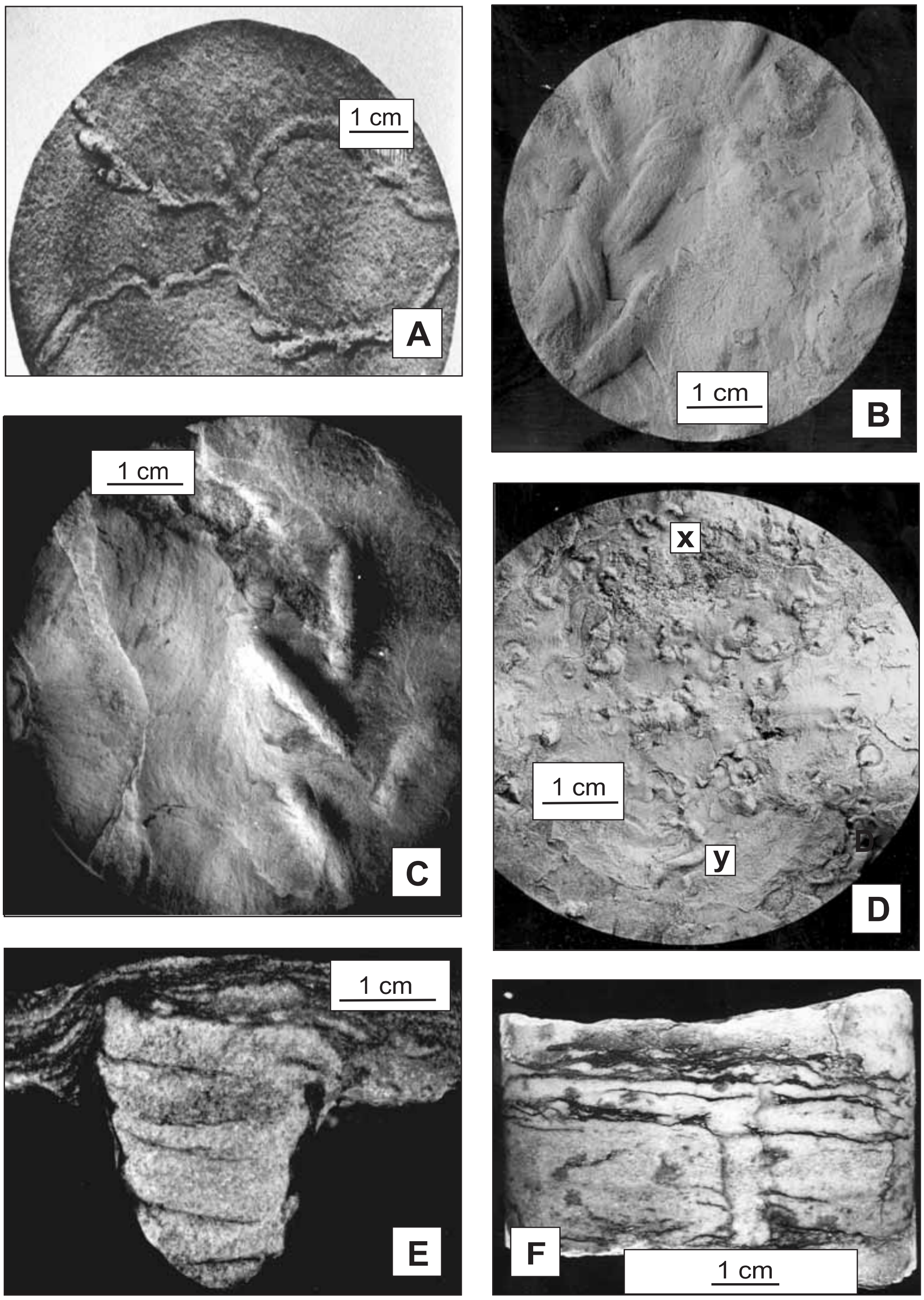

Fig. 36. Trace fossils of the shoreface and offshore zones, early Lower Cambrian, Mazowsze Formation

A – *Trichophycus pedum* (Seilacher), upper offshore facies association, Terebiń IG 5 borehole, depth 3638.2 m; **B** – *Treptichnus triplex* Palij, lower shoreface facies association, Busówno IG 1 borehole, depth 3481.8 m; **C** – *Treptichnus bifurcus* Miller, upper offshore facies association, Łopiennik IG 1 borehole, depth 5223.8 m; **D** – x: *Gyrolithes polonicus* Fedonkin, y: *Planolites montanus* Richter, lower shoreface facies association, Busówno IG borehole, depth 3495.0 m; **E** – *Teichichnus rectus* (Seilacher), vertical cross-section of the retrusive type of feeding burrow, produced by organism moving upwards in order to match the rate of sedimentation, lower shoreface facies association, Radzyń IG 1 borehole, depth 1535.0 m; **F** – *Skolithos* isp., vertical section of the dwelling burrow, lower shoreface facies association, Parczew IG 10 borehole, depth 2102.0 m

storm deposition. The most common is hummocky- and swaly cross-stratification (Raychaudhuri *et al*., 1992). In the analysed material, hummocky cross-stratification can be recorded in sharp-based and scoured sandstone beds of facies *SFh*, *SFL* and *Sx*. However, the limited observation area of drill cores makes it impossible to confidently ascertain the presence of typical storm structures in the analysed sections. In the lower shoreface numerous, thin mudstone interbeds of facies *F* contain abundant trace fossils represented by a diversified assemblage of deposit-feeder burrows of *Planolites mon-*

tanus Richter, *Planolites beverleyensis* (Billings), *Teichichnus rectus* (Seilacher) (Fig. 36E), *Treptichnus triplex* Palij (Fig. 36B), *Gyrolithes polonicus* Fedonkin (Fig. 36Dx) and *Trichophycus pedum* (Seilacher) (Fig. 36A). Filter-feeders' domichnia are represented by rare *Bergaueria* isp. and *Monocraterion* isp. Another domichnia *Skolithos* isp. (Fig. 36F) is more rarely observed.

Upper offshore facies association

Description. The upper offshore facies association is represented by the characteristic domination of facies *SFr*, *SFf* and *SFo*, heterolithic facies *FS* and mudstone facies *F*. Facies *SFp* is very occasional. Another peculiar feature is the occurrence of an ichnotaxonomically and ethologically diversified trace fossil assemblage. The most common are feeding-dwelling burrows (fodinichnia); dwelling burrows (domichnia) are less frequent. Abundance of trace fossils is moderate to high. The bioturbation index (BI) is 4–6. Ichnodiversity is moderate.

Interpretation. The upper offshore zone is situated within maximum (storm) wave base (Howard, Frey, 1984; Pemberton *et al*., 1992b) (Fig. 9). This zone is affected by great storm-wave action only. Flaser, wave and ripple cross-lamination are a relict of storm wave ripples. Most of trace fossils belong to deposit-feeders' fodinichnia represented primarily by *Teichichnus rectus* (Seilacher), *Teichichnus* isp., *Trichophycus pedum* (Seilacher) (Fig. 36A), *Trichophycus* isp., *Planolites beverleyensis* (Billings), *Planolites montanus* Richter, *Planolites* isp., *Treptichnus bifurcus* Miller (Fig. 36C), *Gyrolithes polonicus* Fedonkin and *Neonereites uniserialis* Seilacher. Burrows of deposit-feeders predominate in heteroliths of facies *FS* and mudstone facies *F*, indicating slow mud sedimentation from suspension, which favoured both a high concentration of nutrients in the sediment and its sufficient oxygenation (Fillion, Pickerill, 1990). The predominance of fodinichnia deposit-feeders in low-energy upper offshore environments supports the opinion that increased nutrient contents in the sediment with simultaneous good oxygenation of the basin resulted in higher activity of deposit-feeders. Due to their activity, the bioturbation index in the upper offshore association is high (4–6). The frequency of filter-feeders' dwelling burrows is small. Domichnia are represented by *Bergaueria* isp. burrows and rare *Skolithos linearis* Haldemann and *Monocraterion* isp. They are observed mainly in facies *SFr*. The rarity of filter-feeders' dwelling burrows in the upper offshore as compared with shoreface settings suggests a deepening of the sedimentary environment and its lower energy. The tracemakers favoured higher energy habitats providing more nutrient resources in suspension. Very rare are crawling traces (repichnia) *Gordia* isp.

DEVELOPMENT OF THE ALLUVIAL DEPOSITIONAL SYSTEM

Alluvial depositional systems of the eastern and north-eastern parts of the Lublin–Podlasie basin overlie the late Ediacaran volcanigenic rocks of the Sławatycze Formation. The top of Sławatycze Formation is dated by U/Pb SHRIMP at 551 ± 4 Ma (Compston *et al*., 1995) (Fig. 5). In the Lublin–Podlasie basin, deposition of the Siemiatycze Formation alluvial deposits started probably after a period of epigenetic erosion. Erosional processes removed a large volume of effusive, volcaniclastic and epiclastic rocks. Prior to the erosional processes, volcanigenic deposits were widespread in the western marginal zone of the late Ediacaran trap basalt province.

Sedimentation of the post-volcanic alluvial system is strictly associated with the late Ediacaran tectonic evolution of the Lublin–Podlasie basin. Systems of braided rivers/streams and alluvial fans developed during the late stages of synrift phase. Initially, the deposition took place on topographically

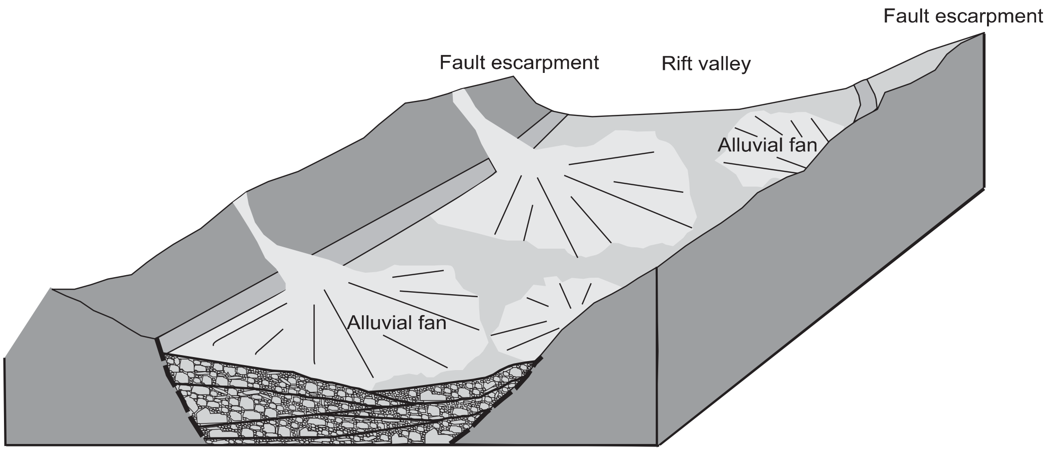

Fig. 37. Rift valley with transverse drainage of alluvial fans. Early stage of alluvial depositional system development in tectonic grabens in the Podlasie part of Lublin–Podlasie sedimentary basin

variable areas in tectonic grabens and half-grabens (Pacześna, 2006). The NW–SE orientation of the axes of large grabens was consistent with the trend of the major rift system that developed during the late Ediacaran on the western margin of Baltica (Poprawa, Pacześna, 2002; Jaworowski, Sikorska, 2003; Pacześna, Poprawa, 2005a, b; Pease *et al*., 2008).

The early stage of the development of the late Ediacaran hydrological drainage of the basin involved the formation of transverse drainage in rift valleys. The characteristic depositional element of this stage is the occurrence of alluvial fans that developed at the foot of fault escarpments in tectonic grabens and half-grabens (Fig. 37). Coarse-clastic alluvial fan deposits observed in the basal parts of borehole sections in the north-eastern region of the Podlasie Depression are good examples of such sedimentation. The presence of coarse-clastic material was most likely associated with synsedimentary faulting in the Iwanki Rohozy–Krzyże depocentre (Pacześna, 2006) (Fig. 38). The coarsest, conglomeratic alluvial fan deposits occur in this area in proximal zones near fault escarpments (near the Iwanki Rohozy section, north of depocentre), and probably represent gravity debris flows. The characteristic features of these conglomerates are grain angular-

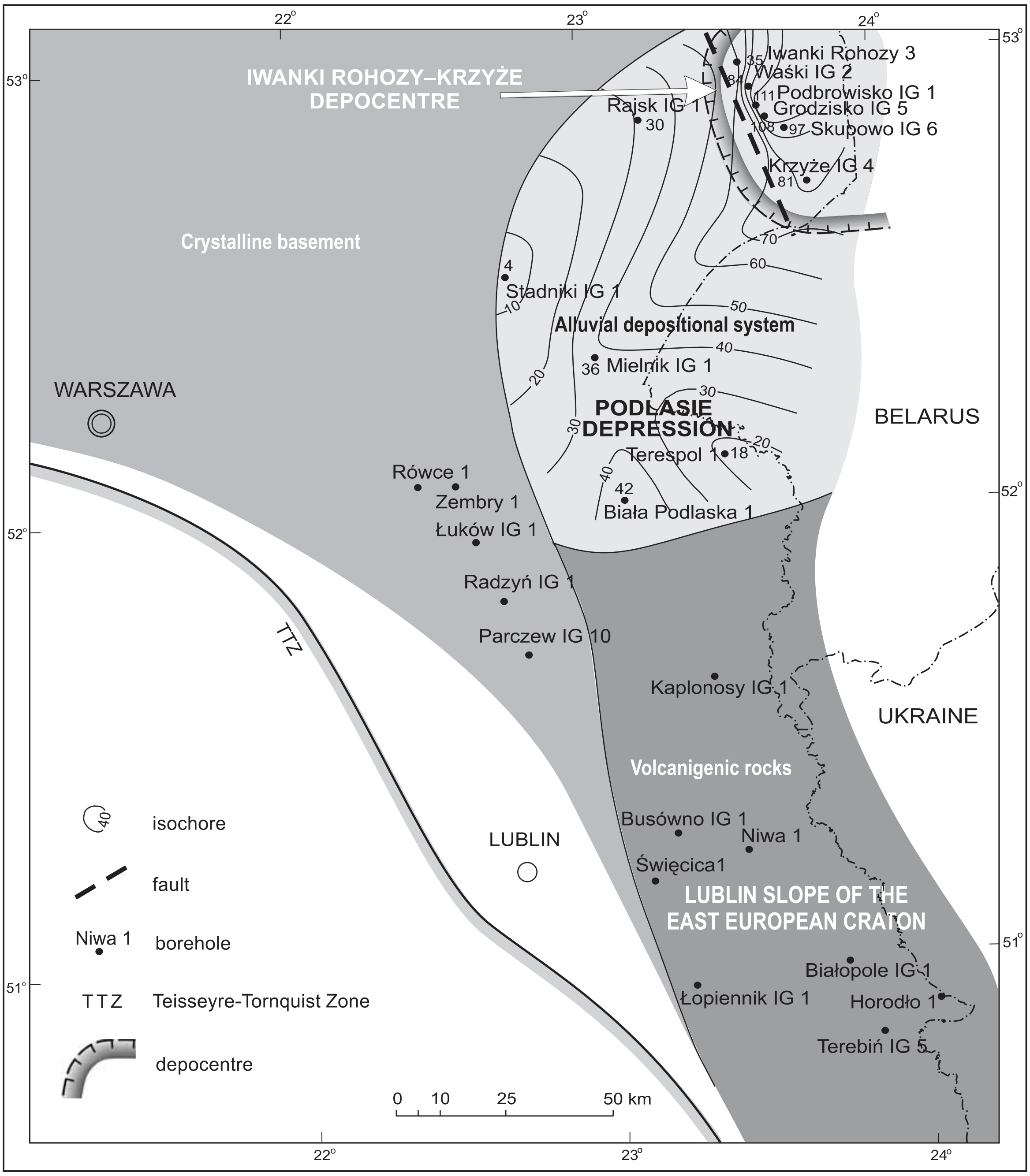

Fig. 38. Combined facies map and isochore map showing location of the Iwanki Rohozy–Krzyże depocentre in the early stage of development of alluvial depositional system (after Pacześna, 2006, modified)

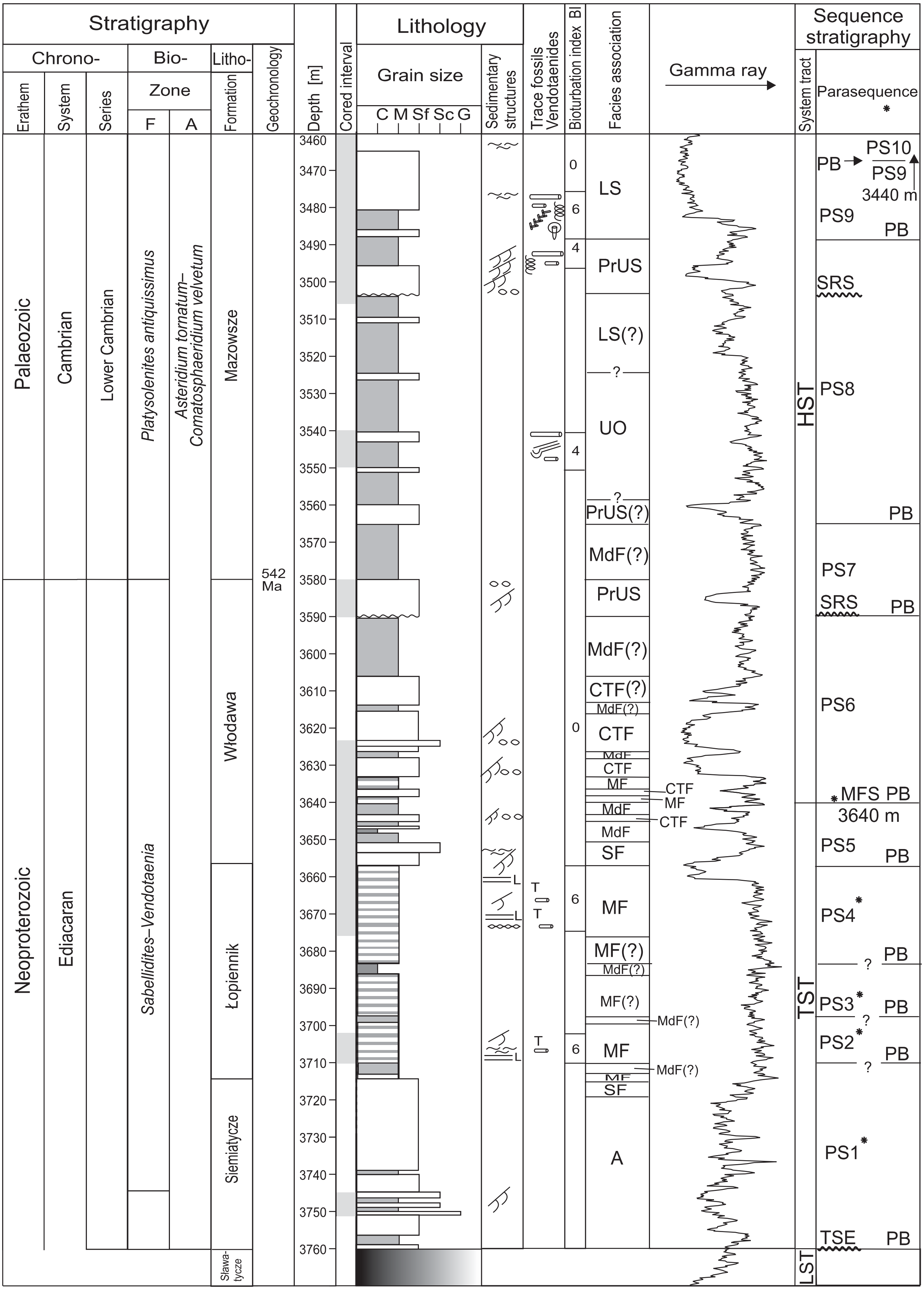

Fig. 39. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Cambrian, Busówno IG 1 section

For explanations see Fig. 17

ity, lack of sorting and massive structure. The gravity debris flows may have occurred as catastrophic flash-flooding events.

Towards the south-east, both within the Iwanki Rohozy–Krzyże depocentre and outside, the grain size gradually decreases. Coarse-grained arkosic sandstones and cross-bedded medium-grained sandstones with rare fine-grained varieties are observed in this area. They represent distal parts of small and steep alluvial fans deposited either from sheetfloods or in shallow braided channels that developed on the fan surface.

Alluvial fans also formed in areas extending near sloping elevations of the crystalline basement. In the Parczew IG 10, alluvial fan deposits occur in the lowermost part of the section (Fig. 10B), directly overlying the crystalline basement. They are represented mostly by very coarse-grained arkosic sandstones containing large feldspar grains, and conglomerates composed of angular clasts of quartz and crystalline rocks. Angular clasts are indicative of short transport. The alluvial fan drainage basin of the Parczew region extended in the area composed of crystalline rocks of the Parczew Elevation. The existence of this elevation was proved by geophysical investigations (Pożaryski, Tomczyk, 1993). Detrital material was supplied to the alluvial fan from the elevation covered with a weathering mantle. The Kaplonosy IG 1 section is also located in the distal part of the alluvial fan probably. Coarse-grained deposits observed in Kaplonosy IG 1 were associated with a hypothetical local crystalline basement elevation. Like in Parczew IG 10, the very coarse-grained sandstones are typical of the entire alluvial interval. In contrast to the Krzyże IG 4 and Parczew IG 10 sections, the Kaplonosy alluvial succession is slightly more fine-grained (comp. Fig. 10C and fig. 10A, B). It can suggest a longer distance to the source area represented by an alluvial fan situated at the foot of a hypothetical crystalline basement elevation.

The alluvial fans were source areas for large braided rivers. During the initial stages of hydrological closure of the rift valleys, drainage basins of the alluvial fans and fluvial braided system were limited to the individual valleys (de Almeida *et al*., 2009). As the rifting processes were propagating from the north to the south and their intensity was decreasing, the relief of the north-eastern areas of the basin became gradually levelled. Axial hydrological drainage became an important depositional factor in the basin. It occurred in rift valleys that developed as a result of coalescing of serial half-grabens into large systems. The systems were probably separated from one another by escarpments which were subjected to gradual degradation as the erosional processes intensified and tectonic activity of the escarpments decreased. These phenomena facilitated expansion of river-drainage processes that caused transfer of fluvial sediments towards the neighbouring depocentres (Smith, 1994; Peakall *et al*., 2000), as there were no elevated structural elements separating the basins.

The general palaeogeographic setting of the alluvial basin was relatively unified in its facies pattern during the final stages of alluvial deposition. The marginal, western and northern regions of the basin were the areas of deposition in distal parts of alluvial fans. These were source areas to large braided rivers flowing down transverse to the axis of the sedimentary basin. The facies record of depositional processes of transverse braided river systems is provided by the Krzyże IG 4 (Fig. 10A), Kaplonosy IG 1 (Fig. 10C), Parczew IG 10 (Fig. 10B) and Busówno IG 1 (Figs. 12C, 39) sections.

At the end of alluvial deposition, the central, axial zone of the elongated alluvial basin was occupied by a SE-sloping alluvial plain covered with an anastomosed system rivers (Fig. 40). Their sedimentary records can be found in the Niwa 1 (Figs. 12B, 41), and Święcica 1 (Figs. 12A, 42) located in the south of the central region of the sedimentary basin. Anastomosed river deposits were also found in Radzyń IG 1 (western part of the basin) (Fig. 15A) and in Mielnik IG 1 (north-eastern part of the basin) (Fig. 15B).

An increase in regional subsidence was an important tectonic process that controlled the facies pattern of the alluvial basin infill. The process caused faster sinking of the southeastern areas of the basin, resulting in a simultaneous increase of

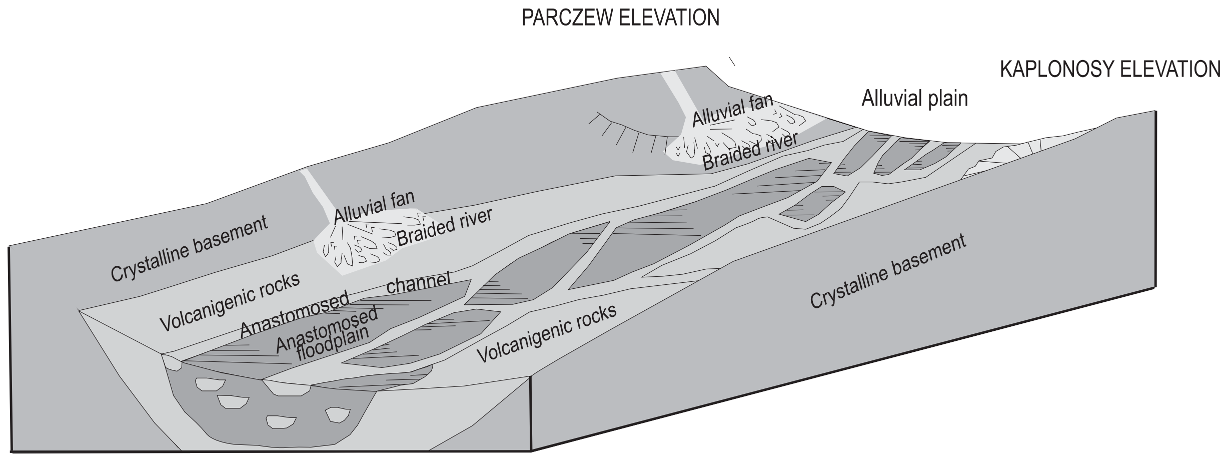

Fig. 40. Late stage of alluvial depositional system development in a rift valley gradually filling with axial anastomosed system rivers deposits. Note occurrence of small alluvial fans and braided rivers near fault escarpments

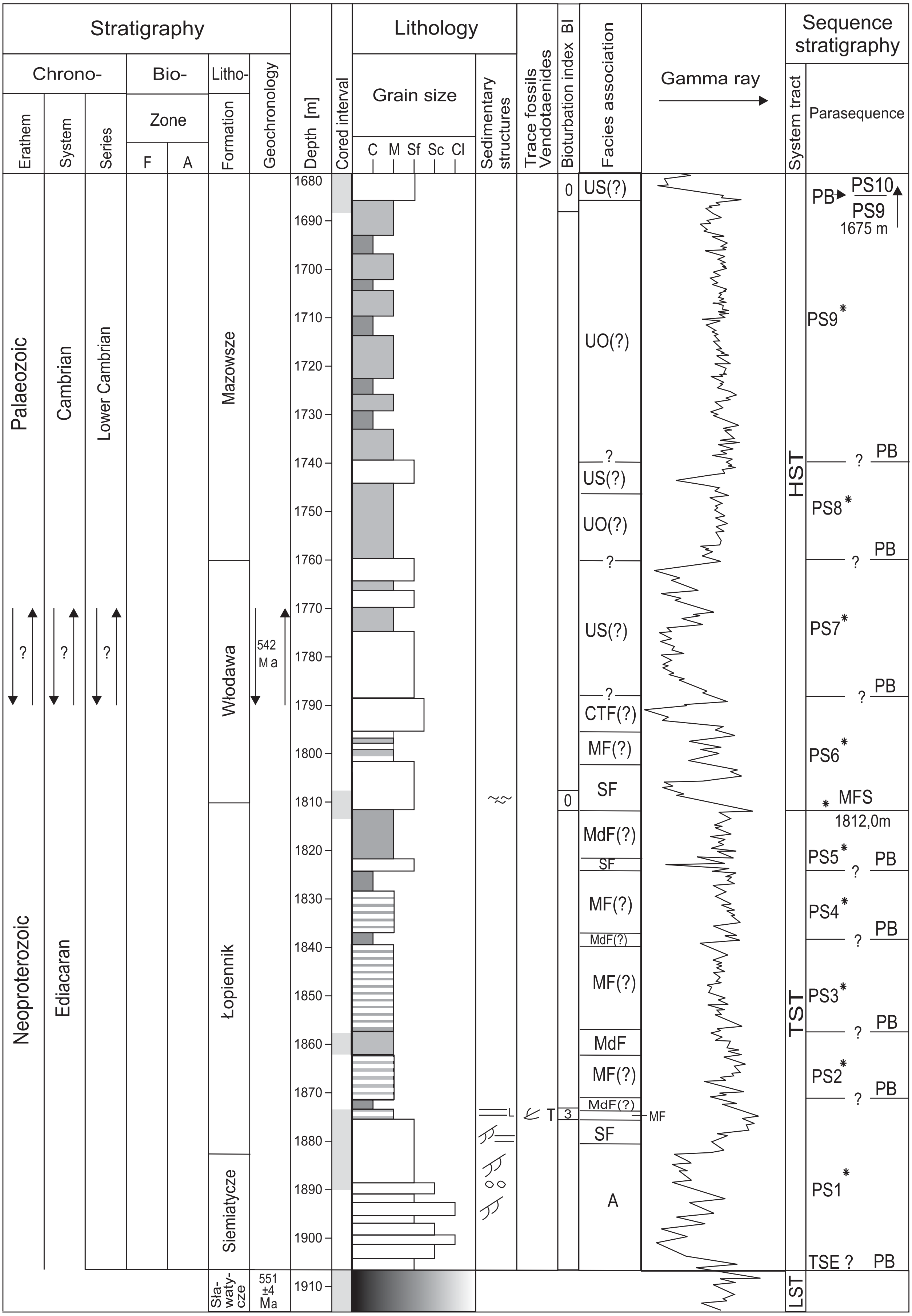

Fig. 41. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Cambrian, Niwa 1 section

For explanations see Fig. 17

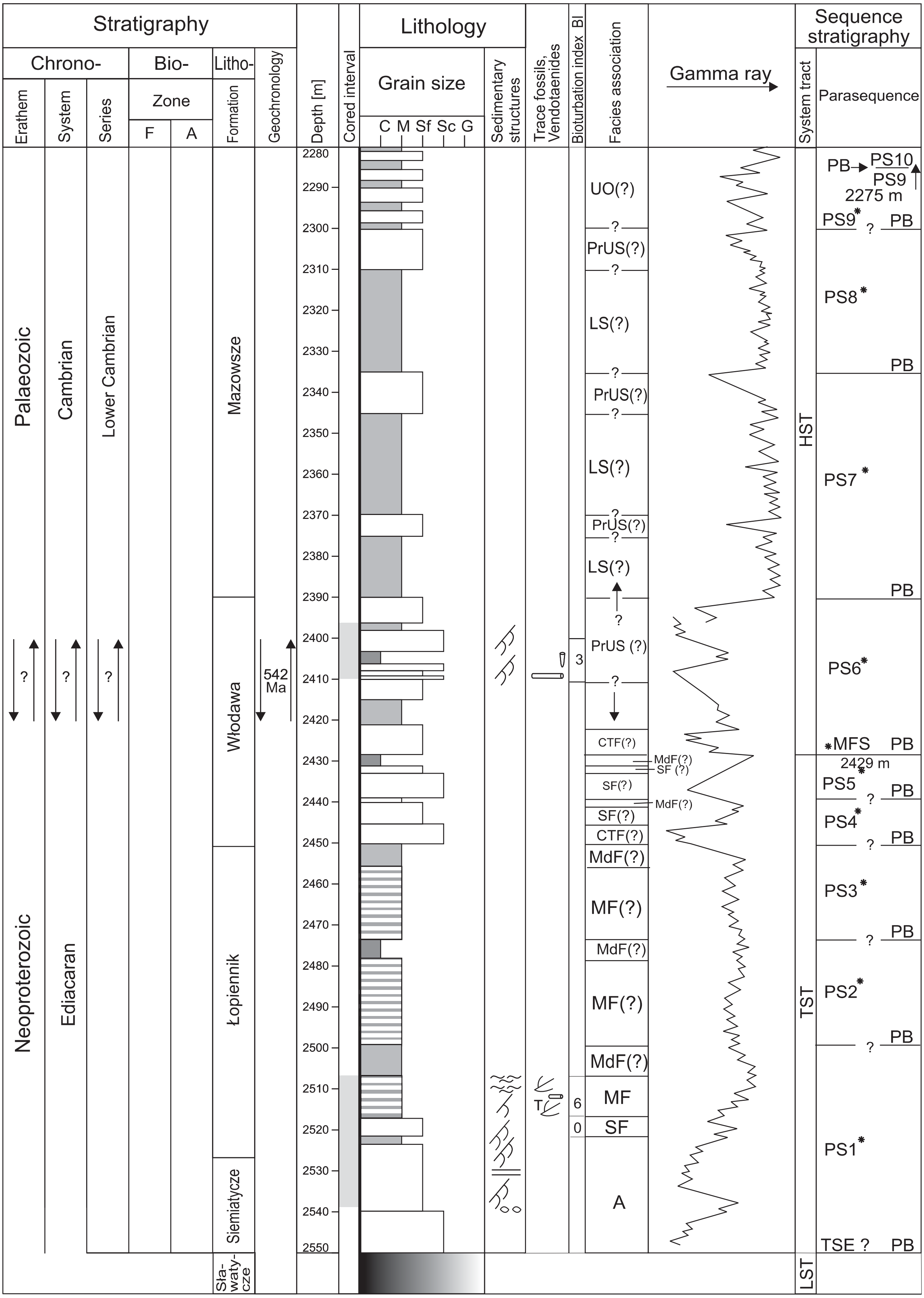

Fig. 42. Grain size, stratigraphy, sedimentary structures, trace fossils, facies associations and sequence stratigraphy, late Ediacaran and early Cambrian, Święcica 1 section

For explanations see Fig. 17

accommodation space and rise of relative sea level. Due to tectonic processes, the area of continental deposition in the basin extended southwards. It was accompanied by a gradual reduction in detrital material supply to the basin because of weaker erosional processes and denudation in source areas. In the facies pattern of the sections, the process manifested itself by the increase in the proportion of fine-clastic deposits represented by medium- and fine-grained sandstones and mudstones in braided floodplain facies successions (Fig. 14A–C) and by the development of anastomosed system rivers.

LATE EDIACARAN CLIMATIC CHANGES – A RECORD IN FACIES

In the western peripheries of Baltica, a long period of an arid and desert climate was gradually replaced by a more humid one. Basinwide alluvial deposition became replaced with time by a variety of very shallow marine depositional settings. The increase in humidity was associated probably with a global sea-level rise after the retreat of the last Gaskiers ice-sheets covering the continents (e.g. Young, 2004; Halverson *et al*., 2005; Fairchild, Kennedy, 2007). The deglaciation process induced a continuous trend of relative sea-level rise in this region of Baltica, lasting until the end of Ediacaran. The climatic changes gave rise to a complete restructuring of the facies pattern in the Lublin–Podlasie basin.

The supposed climatic changes in the western peripheries of Baltica are recorded in the facies of fluvial sections of the analysed basin. The lower part of the sections is a record of deposition in braided rivers and streams associated with alluvial fans in rift valleys. The facies spectrum was dominated by coarse-grained deposits suggesting upper flow regime. During the initial stages, it was a transverse drainage of valleys (Fig. 37). This stage was characterised by the occurrence of episodic, ephemeral braided rivers and streams with very intense flows of upper flow regime. Strong traction and saltation processes occurred in ephemeral river channels. These were typical bedload channels. Both in the past and today, they occur in an arid and semi-arid climate. Ephemeral discharge was probably controlled by seasonal, low annual rainfall (Hiller *et al*., 2007). Under such climatic conditions, very intense weathering processes took place in source areas and resulted in a supply of large amounts of coarse-grained material to the ephemeral braided river basins.

Upper in the braided river sections, a fining-upward trend is observed both in channel and floodplain deposits (Fig. 10B, 14A–C). Although a lot of sediment was still transported in upper flow regime, the increase of fine-grained facies suggesting lower flow regime can be related to a change in hydrological regime.

The fining-upward trend in deposits suggesting lower flow regime of the fluvial deposits can reflect braided perennial or semi-perennial rivers with episodes of intense floods (e.g. Long, 2006). The braided perennial rivers existed under a more temperate climate as compared with climatic conditions under which the lower parts of the sections were accumulated. Increased humidity is proved by more common deposition in crevasse splays in anastomosed system rivers (Fig. 15A), associated with strong flood surges.

Another fact, which can indicate different climatic conditions between the sections, is a change in colour of the sediment. The lower part of the alluvial sections shows brownish and red to variegated colouration of coarse- and fine--grained deposits. Upper in the sections, there is a gradual transition into grey and yellow colour in sandstones, and into black and dark grey colour in claystones and mudstones. Brownish colouration of deposits is due to the presence of large amounts of iron compounds in the matrix (cement), originating from chemical and mechanical weathering in a hot and arid climate (deserts and semi-deserts). Dark colours of continental deposits are typical of sedimentation in a more humid climate.

The alluvial-estuarine succession covers the time interval from ca. 551 ±4 Ma to ca. 542 Ma. The former date is constrained by U-Pb zircon age of the uppermost tuff horizons that underlie the aluvial deposits (Compston *et al*., 1995). The latter defines geochronological age of the Ediacaran/Cambrian boundary (International Commission on Stratigraphy, 2010) and roughly corresponds to the age of the top of the succession under study. Taking into account the relatively long time interval of the succession (ca. 9 Ma) it can be hypothetically assumed that the potential climatic changes recorded in the alluvial system facies could be related to the drift of Baltica after the final stages of breakup of the Rodinia supercontinent. Palaeomagnetic data indicates that Baltica and nearby Laurentia drifted from the moderate southern latitudes (60–30°S) (Nawrocki *et al*., 2004; McCausland *et al*., 2007) or alternatively moderate northern latitudes (e.g. Popov *et al*., 2002;) to the palaeoequator during the period of about 560–540 Ma. During the drift, Baltica was rotated anticlockwise of ca. 120°. Palaeomagnetic data, suggesting the proximity of these two continents, are supported by a similarity between microbiota from different sections of Baltica and Laurentia. Taxonomically similar assemblages of cyanobacteria and acritarchs suggest shallow-marine shelf areas extending between Baltica and Laurentia, which enabled migration of benthic microbiota and phytoplankton (Moczydłowska, 1991; Pease *et al*., 2008). The ultimate separation of the continents occurred about 560–550 Ma (e.g. Cawood *et al*., 2007; Elming *et al*., 2007).

The climatic changes may have been related to the drift of Baltica from middle southern latitudes to equator ca. 560–540 Ma. As at ca. 547–548 Ma, Baltica was in the palaeoequator zone and the marine transgression progressed, climate became more humid. The late Ediacaran restructuring of the facies pattern may have been caused by climatic changes. Both the progressing marine transgression and more temperate climate gave rise to the formation of two facies zones in the late Ediacaran basin. The northern and northeastern (Podlasie Depression) zone of the basin remained continental and characterised by fluvial sedimentation in braided rivers with a fluctuating but perennial discharge. In the central and southeastern area, alluvial sediments gradually passed into a variety of marine deposits of tidal depositional environments.

FACIES ZONATION OF THE LUBLIN ESTUARY

FLUVIAL-MARINE TRANSITION – INNER ESTUARY

One of the arguments for estuarine origin of the late Ediacaran siliciclastic succession in the Lublin–Podlasie basin was the occurrence of the characteristic tripartite vertical succession in the north of the Lublin region: alluvial deposits – deposits of alluvial/tidal transition-tidal deposits. The succession is observed in the Kaplonosy IG 1 (Fig. 20), Parczew IG 10 (Fig. 19) and Radzyń IG 1 (Fig. 17) sections. The middle member of the succession corresponds to a zone where facies associations representing continental fluvial deposits are mixed with subordinate tidal deposits.

According to the classical definition proposed by Dalrymple *et al*. (1992), an estuary is *the seaward part of a drowned valley system which receives sediment from both fluvial and marine sources and which contains facies influenced by tide, wave and fluvial processes. The estuary is considered to extend from the landward limit of tidal facies at its head to the seaward limit of coastal facies at its mouth*.

The definition determines development of the estuary as dependent on the relative sea-level rise during a transgression. According to the above presented definition, another argument that can support the occurrence of estuarine deposits in the succession is its transgressive nature controlled by local tectonic factors (Pacześna, Poprawa, 2005a, b).

Referring to the terminology and facies zonation in estuaries (Dalrymple *et al*., 1992), the fluvial-marine transition zone represents the inner estuary referred to as the upper estuary by some authors (e.g. Borrego *et al*., 1995; Carr *et al*., 2003; Greb, Martino, 2005).

The lower member of the succession from the northern area of the Lublin region is composed of alluvial deposits representing alluvial fans and braided rivers and streams. They were discussed in detail in the previous chapters. Therefore, only the topmost parts of the alluvial succession will be discussed here briefly. They are directly overlain by fluvial-marine deposits of the transition zone.

The fluvial-marine transition zone from Kaplonosy IG 1 is the best marked among the three sections. It was encountered at a depth of 1423.0–1401.5 m (21.5 m in thickness). The following facies succession is observed (Fig. 43A):

> *SCl*→*SCm*→*SCx* (braided river channel) →*SFm*→*SCx*→*SFm*→*SCx*→*SFm*→*SCx*→*SFm*→*SCx*→*SMx*→*SCx*→*SMx* (braided floodplain) →*SMx*→*SMl*→*SFr* (transition interval–braided river channel/tidal channel) →*Fh* (mixed tidal flat) →*SCl* (tidal channel) →*Fh* (mixed tidal flat).

In Parczew IG 10, the transition zone is recorded at a depth of 2283.0–2271.5 m (11.5 m in thickness). The facies succession is as follows (Fig. 43B):

> *SCl*→*SM* (braided river channel) →*SFm* (braided floodplain) →*SC* (braided river channel) →*SFm*→*SCx*→*SFm* →*SCx*→*F* (braided floodplain) →*SCx* (transition interval–braided river channel/tidal channel) →*SFr*→*SFf* (sand tidal flat) →*SCm* (tidal channel) →*Fh* (mixed tidal flat).

The Radzyń IG 1 section differs from the remaining two sections in the development of the transitional succession. It was recorded at a depth of 1657.0–1640.6 m (16.4 m in thickness). Channel deposits of facies *SCm*, 2 m thick, and the 8 m thick package of floodplain deposits with crevase splays, composed of interbedding mudstone facies *F* and sandstone facies *SFm* is overlain by tidal channel deposits represented by facies *SCx* and *SCr*. The following members are observed in the transitional succession (Fig. 43C):

> *SCm* (channel) →*F*→*SFm*→*F*→*SFm*→*F*→*SFm*→*F*→*SFm* →*F*→*SFm*→*F*→*SFm*→*F*→*SFm*→*F*→*SFm*→*F*→*SFm* →*F*→*SFm*→*F*→*SFm*→*F* (anastomosed river floodplain with crevasse splays) →*SCx*→*SCr* (transition interval–tidal channel/anastomosed river channel) →*Fh* (mixed tidal flat) →*SCx* (tidal channel) →*Fh* (mixed tidal flat).

There are few direct evidences of tidal activity in the transition zone. In each of the three sections, the uppermost fluvial package represented by both channel and overbank deposits is directly overlain by sediments which are only slightly modified by tidal currents.They consist of coarse-grained facies *SC*, *SCp*, *SCl* and *SMr* and thinly laminated sandstone-mudstone-claystone heteroliths of facies *Fh*. The last facies represents tidal rhythmites deposited on the mixed tidal flat. Thicknesses of the first packages of heteroliths vary between 0.30 and 0.50 m in the Kaplonosy IG 1 and Radzyń IG 1 sections. In the Parczew IG 10, tidal channel deposits are overlain by fine-grained sandstones of facies *SFf*, representing the sandflat. Although rhythmites are deposits typical also of fluvial terraces and lake littorals, their tidal origin is confirmed here by the co-occurrence with other tidal indicators such as mud diapirs and reactivation surfaces. A marine origin of heterolithic deposits is definitely supported by frequent trace fossils of *Planolites montanus* Richter and *Torrowan-*

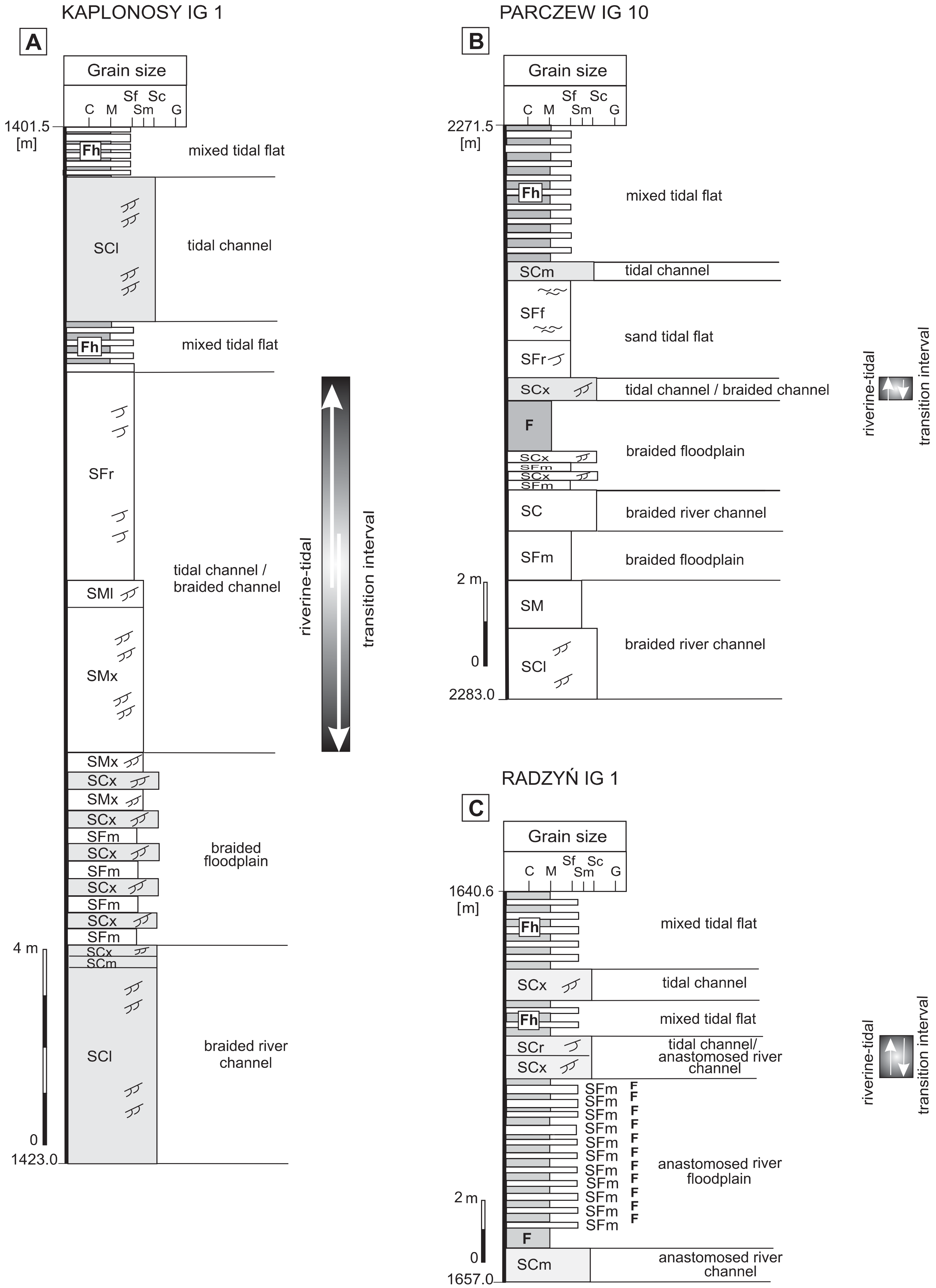

Fig. 43. Facies successions of the transition zone representing the upper part of the Lublin estuary. Note occurrence of mixed tidal channel – braided channel deposits in the Kaplonosy IG 1 and Parczew IG 10 and mixed tidal channel – anastomosed river channel deposits in the Radzyń IG 1

For explanations of grain size, sedimentary structures and facies code see Fig. 10

gea rosei Webby as well as Cyanophyta thallus of *Vendotaenia antiqua* Gnilovskaya.

Other evidences of tidal origin of heterolithic deposits are thin (often up to 1 mm in thickness) mud drapes occuring within cross-beds of sandstone facies *SFl* and rarely *SMl*. The sandstones form thin, up to 30 cm thick beds in sandstone-mudstone-claystone heteroliths. The main reason for small thickness of mud drapes was a low content of mud material in the transition zone. It results from the fact that it was the zone of strong activity of river currents that transported mainly coarse-grained material into the sea. Small amounts of mud material were deposited in the transition zone by tidal currents on inter-channel mud shoals. Thickness of these sparsely represented deposits does not exceed 0.5 m.

Direct evidences for the existence of tides in the upper estuary area include also occurences of bimodal cross bedding as a record of current reversals. Bimodal cross bedding was identified in coarse-grained sandstones of facies *SCx* and *SCl* in the Kaplonosy IG 1 and Parczew IG 10 sections.

The same facies succession is observed in all the analysed sections: facies *Fh* appears as the first, being underlain by facies *SCr*, *SCx*, *SCl* and *SMr*. Direct contact between the tidal flat deposits and the underlying coarse-grained sandstones indicates that the latter could represent tidal flat channels. Genetic relationships between the mixed tidal flat and the tidal channel facies association are suggested by their immediate superposition in the vertical succession. It can be related to the occurrence of tidal flats along the tidal channel banks.

Mixed tidal flat deposits have been often observed in the upper zones of ancient estuaries. However, it is commonly a fragmentary record proving their poor development in the inner estuarine zones (e.g. Hori *et al*., 2001; Carr *et al*., 2003; Dalrymple, Choi, 2007; Kitazawa, 2007; Hovikoski *et al*., 2008). In the fluvial-marine transition zone of modern estuaries, mixed tidal flat deposits commonly occur in the upper parts of inclined heterolithic stratification of channel margin successions (Dalrymple, Choi, 2007).

Indirect evidences for tidal deposits in the succession include both the occurrence of the tidal facies succession and the underlying packages of low-angle and high-angle cross-bedded coarse-grained sandstones. The latter can suggest both fluvial and tidal depositional environments and suggest respectively upper or lower flow regime and indicate deposition under conditions of increasing or decreasing energy. In each of the three sections, the uppermost fluvial package represented by both channel and overbank deposits is directly overlain by sediments which are only slightly modified by tidal currents.

The relative scarcity of tidal indicators in the fluvial-marine transition zone results from the fact that the effect of tides on both depositional processes and morphology of coarse-grained deposits is limited as compared to the effects of fluvial and wave activities. Tides show low ability to remobilize deposits containing the coarse-grained sand and gravel fractions (Dashtgard, Gingras, 2007).

In the transition zone, a net seaward transport occurred due to relatively strong fluvial flows. The supply of fluvial material increased seasonally during periods of high discharges of the rivers. In the fluvial-marine transition zone, marine and fluvial deposits were mixed. The presence of transition zone could cause development of braided tidal-channel networks in the initial phase of estuary development. The phenomenon of braided character of tidal channel is relatively rare in modern and ancient estuaries (Dalrymple, Choi, 2007). The Ediacaran environmental conditions, especially lack of vegetation and low mud content, resulted in instability of channel banks in the transitional zone (Smith, 1976). Due to these factors, the channels could have poorly defined morphology of a braided character. Because of a steep gradient of braided plain, the zone of mixing of tidal and fluvial processes should be short (Dalrymple *et al*., 1992; Dalrymple, Choi, 2007). Transition zone of the Lublin estuary was probably relatively narrow and extended over at most several kilometres to the south of the line Radzyń IG 1–Parczew IG 10–Kaplonosy IG 1. The late Ediacaran channels from the Lyell Land Group, northeast Greenland, can be an example of such a braided pattern of tidal channels. These are shallow, very wide channels of relatively poorly defined morphology, continously shifted position and cut into previously active channels. This created the architecture of multistorey tidal channel sand sheets (Tirsgaard, 1993).

The presence of braided patterns of tidal channels in the transition zone can be proved by both the occurrence of very coarse-grained deposits of facies *SC* and *SCx* composing tidal channel fills. Assuming the existence of braided patterns of tidal flats in the inner estuary, we can presume that the tidal channel fills may have been represented by transverse sand bars composed of coarse-grained sandstone facies *SC*, *SCx* and *SMl*, like in braided channels. Based on the spectrum of sedimentary structures in the channel fills, can recognise a record of upward decreasing flow energy within the environment. The following facies sequence can prove the existence of such a process: *SMx*→*SMl*→*SFr*→*Fh* (Kaplonosy IG 1; Fig. 43A).

The studied ancient estuary was most likely a hypersynchronous as defined by Dalrymple, Choi (2007). This means that the funnel-shaped geometry of estuary caused an increase in the tidal range and speed because of the progressive decrease in cross-sectional area and increase of convergence of the banks of estuary in transitional zone (Fig. 44). In addition, convergence could be enhanced by a specific relief of areas surrounding the transition zone. The transition zone convergence was constrained by the presence of two, located opposite in a short distance, elevations of the Palaeoproterozoic crystalline basement, represented by the geophysically proved Parczew elevation and the hypothetical Kaplonosy elevation. Hypersynchronous conditions are typical of tidal-dominated estuaries (e.g. Dalrymple *et al*., 1992; Kitazawa, 2007).

Changeability of sediment dynamics caused by cyclic changes in hydrological regime in tidal channels of the transition zone roughly corresponded to the constraints on the deposition of transverse bars in braided channels during both water level drops and decreases in flow energy. Additionally, low proportion of mud and an absence of vegetation make that studied deposits are similar to facies successions from river channels (Eriksson *et al*., 1995).

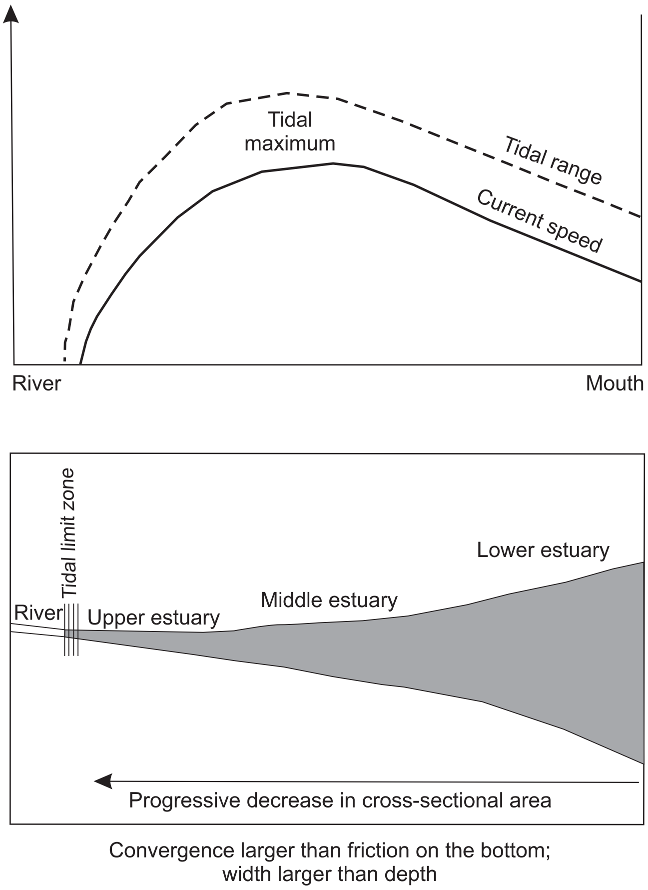

Fig. 44. Hypersynchronous system of tidal range and tidal current speeds along the length of a macrotidal-dominated estuary. The progressive decrease in cross-sectional area in funnel-shaped estuary causes the tide to increase in range. Beyond the tidal maximum point, friction on the bottom causes tidal range and tidal current speeds to decrease to zero at the tidal limit (after Dalrymple, Choi, 2007, modified)

In the area north of the Kaplonosy IG 1–Podedwórze IG 2–Parczew IG 10–Radzyń IG 1 line, there are neither tidal deposits nor signs of effects of tidal currents in the fluvial successions. Alluvial deposits are the only ones in this area (Pacześna, 2006). Therefore, it may be inferred that the tidal limit zone extended over a small area, a few kilometres to the north and south of these boreholes.

Another argument supporting tidal nature of the estuary is the lack of bay-head delta deposits in the transition zone during the transgressive stage of the estuary evolution. They are typical of wave-dominated estuaries and are deposited by rivers at the head of the estuary in a zone of their influence. In tidal estuaries, strong tidal currents push clastic material carried by rivers back into river channels, thus preventing from both deposition and progradation of delta lobes in the upper estuarine zones.

None of the sections from the transition zone shows coarsening upward cycles that could suggest a progradation trend within the succession.

The upper part of both ancient and modern estuaries is characterised by the presence of straight tidal channels. Their sinuosity increases seawards (e.g. Hori *et al*., 2001; Greb, Martino, 2005; Eriksson *et al*., 2006; Kitazawa, 2007). No meandering channels on tidal flats have been reported to date from the Precambrian deposits. There are also no meandering rivers (Eriksson *et al*., 1998, 2006). It is assumed that Precambrian tidal channels were straight or of very low sinuosity. In most cases, these were wide sandy tidal channels filled with a few metres thick sand sheets (Tirsgard, 1993; Corcoran *et al*., 1998; Eriksson *et al*., 2006).

In the estuary conditioned by marine transgression, two main trends revealed within the vertical facies succession of the fluvial-marine transition zone. The first one is a progressive upward increase in the proportion of tidal facies associations and the presence of direct evidence for tides. The facies associations represent deposits of both tidal channel environments and other zones of tidal flats (sandflat, mixed flat and mudflat). A higher frequency of evident tidal indicators such as mud drapes on the foresets within cross-bed sets, reactivation surfaces and cyclic rhythmites proves an increasing influence of tides in creating the facies pattern as the transgression advanced.

The other trend is a remarkable upward decrease in the content of coarse grains in sandstones observed in the transgressive estuarine succession (Fischbein *et al*., 2009). This is also a record of decreasing fluvial effect on sedimentary processes in the transition zone. The probable record of fluvial effects is manifested only by the presence of coarse-grained facies in tidal channel fills in the lower tidal part of the succession. These represent channels that developed mainly on the sandflat, rarely on the mixed flat and mudflat. In the facies spectrum, the channel fills are represented mostly by facies *SCl*, *SCm* and *SCx*. Towards the top of studied succession, the thickness of channel fills increases. This fact can suggest increasing depths of the channels. The shallowest channels occurred in the lower part of the tidal succession. Their inferred depth did not exceed 1.0 m. As the effect of tides was increasing in the basin, the channels were getting much deeper. Their depth in the upper part of the tidal succession of the Radzyń section is estimated 3.5–4.5 m. It is worth noting that most of the deep channels were associated with the sandflat. Tidal channels of other were shallower and commonly developed on the mixed flat.

The inner estuary is featured by very well developed sandflats. Thickness of the sandflat deposits varies from 1.0 to 3.5 m. The sandstone packages consist of facies *SCl*, *SFh*, *SFl* and *SFm*. The fine-grained sandstones are typically well sorted. Additionally, facies *SFr* and *SFf* are observed in some sandflat packages. Facies *SFh* occurs most frequently within the sandflat facies association. A good example of the fact is the sandflat deposits observed in the Podedwórze IG 2 section (Fig. 18). Horizontal stratification indicates a deposition in extremely high-energy tidal flat environment of upper flow regime. This is a zone of high velocity tidal currents characteristic of the tidal-energy maximum in tide-dominated estuaries (Dalrymple *et al*., 1990; Greb, Martino, 2005; Kitazawa,

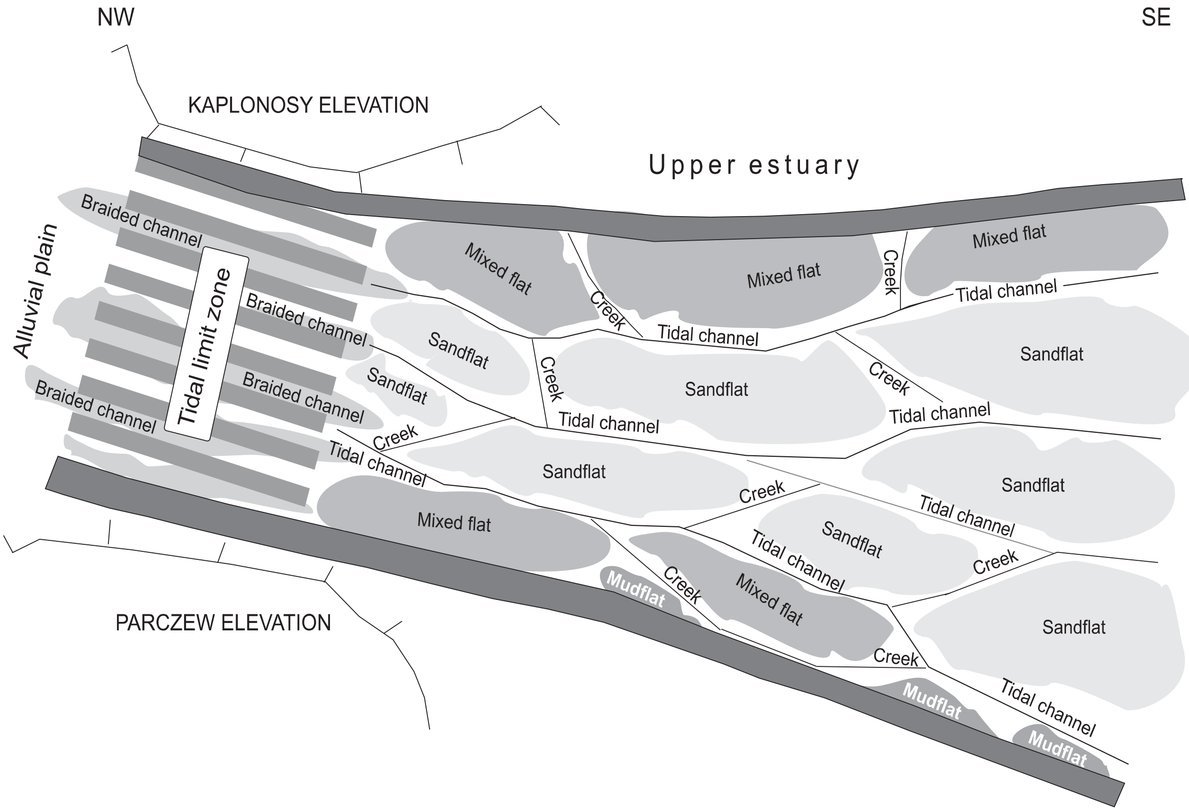

Fig. 45. Braided pattern of tidal flat in the upper Lublin estuary during the earliest stage of parasequence PS1 development

2007). It seems problematic whether the upper-flow regime tidal flat occurs in the inner estuary. High-energy sandflats of ancient and modern tide-dominated estuaries are situated within the middle estuary. In the Lublin estuary, the location of high-energy sandflats in the upper estuarine parts can probably be explained by the braided pattern of tidal flats (Fig. 45). High energy of the deposition on braided sandflats could completely prevent meandering tidal channels development. However, the braided sandflats were characterised by straight channels in both the upper and middle estuary. In the tide-dominated Lublin estuary, with an extensive zone of braided sandflats that expanded seawards during transgressive events, the low-energy zone may have been absent.

MIDDLE ESTUARY

All tide-dominated systems (estuaries and deltas) contain a mixed-energy, fine-grained sand dominated bedload convergence zone that is situated below tidal limit zone, between an inner, fluvially dominated part of the estuary that has net seaward-directed transport and an outer, tide-dominated part that has landward-directed transport. In modern estuaries with small river, bedload convergence occurs a few kilometres away from the tidal limit zone, within tidal-fluvial transition. An example is the Cobequid Bay-Salmon River estuary, eastern coast of Canada (Dalrymple *et al*., 1991). In estuary with large river, the Gironde estuary, bedload convergence is situated ca. 20 km from tidal limit zone and ca. 30 km from the estuary mouth, in middle part of estuary (Allen, 1991; Dalrymple, Choi, 2007). The middle estuary extends between of bedload convergence zone and a region located near the mouth of estuary (Dalrymple, Choi, 2007).

The facies spectrum of the Busówno IG 1, Niwa 1 and Święcica 1 sections, located in the south of the central region of the Lublin–Podlasie basin (Fig. 2), is dominated by heterolithic facies *Fh* and fine-grained sandstone facies *SFx*, *SFf*, *SFr*, *SFo*. The proportions of mudstone and claystone facies *F* also remarkably increased. Compared with the upper estuary, the proportion of fine-grained facies in the sections increased by an average of 40% for sandstone facies (comp. Fig. 17 and Fig. 37). A particularly remarkable increase by almost 60% is observed in the proportion of finely laminated sandstone–mudstone–claystone heteroliths of facies *Fh* (comp. Fig. 17 and Fig. 39).

No coarse-grained facies are observed in the transgressive succession of these sections (Figs. 39, 41, 42). Coarse-grained sandstones of facies *SCx* compose two 1 m and 2 m thick beds only in the Busówno IG 1 section. Towards the southeast, there is a gradual increase in grain size. The domination of fine-grained fraction suggests that there was a bedload convergence zone in the south of the central region of the basin, where the estuary's finest-grained deposits occurred. In the Lublin estuary, the probable location of bedload convergence in the central area of the basin ca. 30 km from tidal limit zone and 60 km from mouth of estuary can suggest the upper part of the middle estuary.

Compared with the inner estuary, the typical sedimentary feature of middle estuarine deposits is characterized by considerably increased frequency of direct indicators of tide occurrence. In the Lublin estuary, they are observed mainly in tidal flat and tidal channel deposits. Especially conspicuous is the increased number of mud drapes. They form a thin layers (commonly up to 5.0 mm thick and locally discontinuous) overlying ripple-and cross-bedding foresets in fine-grained sandstone beds that occur between the packages of sandstone-mudstone-claystone heteroliths (Fig. 33A). The frequence and thickness of mud drapes increase towards the southeast. In the upper estuary (Parczew IG 10, Radzyń IG 1 and Podedwórze IG 2) they are sporadic. The frequency of mud drapes for every 1 m of sedimentary logs is 2–4. In the central part of the estuary (Busówno IG 1, Niwa 1, Święcica 1), mud drapes are more frequent: 4–8 mud drapes occur per 1 m of the sedimentary log. In the southeasternmost part of the estuary (Białopole IG 1, Łopiennik IG 1, Horodło 1 and Terebiń IG 5), 1 m of the sedimentary log contains 10–14 mud drapes.

When the river enters the brackish-water area in estuary, suspended silt and clay sized particles begin to form aggregates called flocs in responce to the electrical reaction between the ions and a complex of organic molecules in water. This is a process called flocculation and occurs where salinity is in the range of 1–10‰. An elevated zone of fluid mud accumulation associated with high flocculation rates is a turbidyty maximum and is a phenomena that has been recognized in most modern estuaries as a result of vertical circulation patterns in an estuary due to the landward flow of tidal waters and the seaward flow of fresh fluvial waters (Dörjes, Howard, 1975; Ranger, Pemberton, 1992). This zone is characterised by turbidities and suspended sediment concentrations greater than that occur upstream towards the fluvial system of the upper estuary, or downstream towards the lower estuary. Turbidity maximum zone extends between the tidal limit zone in the upper estuary and the zone where the salinity is ca. 20‰. In most estuaries it is the middle part of estuary (Dalrymple, Choi, 2007). The increase in the number of mud drapes in the middle estuary is associated with a turbidity maximum zone. Mud drapes form as a result of deposition and resuspension of flocks by tidal currents during a single slack water period. Changes in suspended-sediment concentration, affecting the frequency of mud drapes, are observed along the middle estuary. Increased content of suspended sediment occurs below the zone of elevated suspended-sediment concentrations (turbidity maximum), causing development of relatively thick mud drapes. This is observed in the outermost sections of the southeastern part of the Lublin estuary: Terebin IG 5 (Fig. 23, interval 3760–3764 m), Łopiennik IG 1 (Fig. 22, interval 5472–5474 m, 5480–5481 m) and Białopole IG 1 (Fig. 25, interval 2904–2929 m, 2916–2918 m).

The middle estuary (e.g. Dalrymple, Choi, 2007) is an area of maximum velocity of tidal currents and the occurrence of upper flow regime structures. The high-velocity tidal current zone is situated within the marine-dominated part of estuary, near its mouth. In the Lublin estuary, it can be represented by sandflats composed of fine-grained sandstone facies *SFh* and, rarely, *SFl*. Upper flow regime sand flat deposits are observed in the Łopiennik IG 1 (Fig. 22) and Terebiń IG 5 (Fig. 23) sections. A package representing sandflat deposits in Terebiń IG 5 is 15 m thick. It consists of fine-grained sandstone facies *SFh* and *SFl*. The presence of horizontal stratification proves both the dominance of bedload transport (Alexander *et al*., 1991) and high-energy conditions of tidal currents. Bimodal planar cross-bedding is also observed here. It is a record of current reversals herringbone cross-bedding and a good indicator of tides, especially when supported by other tidal indicators in the succession (Eriksson *et al*., 1981).

The tidal succession of the middle part of the Lublin estuary is dominated by widespread deposits of the intertidal zone. They account for approximately 70–85% of the entire transgressive tidal succession interval of each section in this part of the estuary.

Middle estuary sandflat is the least common association of the intertidal zone. Most of the packages are commonly thin, ranging between 0.3 and 4.0 m. There are only two packages found in Terebiń IG 5 and Święcica 1, which attain thicknesses of 15 m and 14 m, respectively. The thickness of the sandflat packages decreases upwards in the intertidal zone succession.

The middle part of the intertidal succession, corresponding to the mixed flat, is occupied by a thick complex of characteristic very finely laminated sandstone–mudstone–claystone heteroliths. Planar laminated tidal rhythmites are characterised by alternating bedding consisting of very thin layers of fine-grained sandstone and mudstone or claystone (Fig. 46A, B). The rhythmites reveal changes in laminae thickness, reflecting variability in the amount of sand and mud material transported and deposited by tidal currents. The above variability is a response to cyclic changes of velocity of tides. The sandstone and mudstone–claystone laminae form the couplets. Each couplet, 1 to 8 mm thick, is composed of a sandstone lamina overlain by a mudstone or claystone one. Thicker sandstone laminae show very small-scale ripple structures often of opposed climb-directions. They indicate that the deposition occurred during a few tidal cycles with intervening short slack-water episodes (Porębski, 1995). Ripple cross--lamination is commonly represented by cosets up to 0.5 cm thick. Lenticular and flaser lamination is less frequent.

The vertical succession of tidal rhythmites contains tidal bundles consisting of 10–13 couplets. They are separated

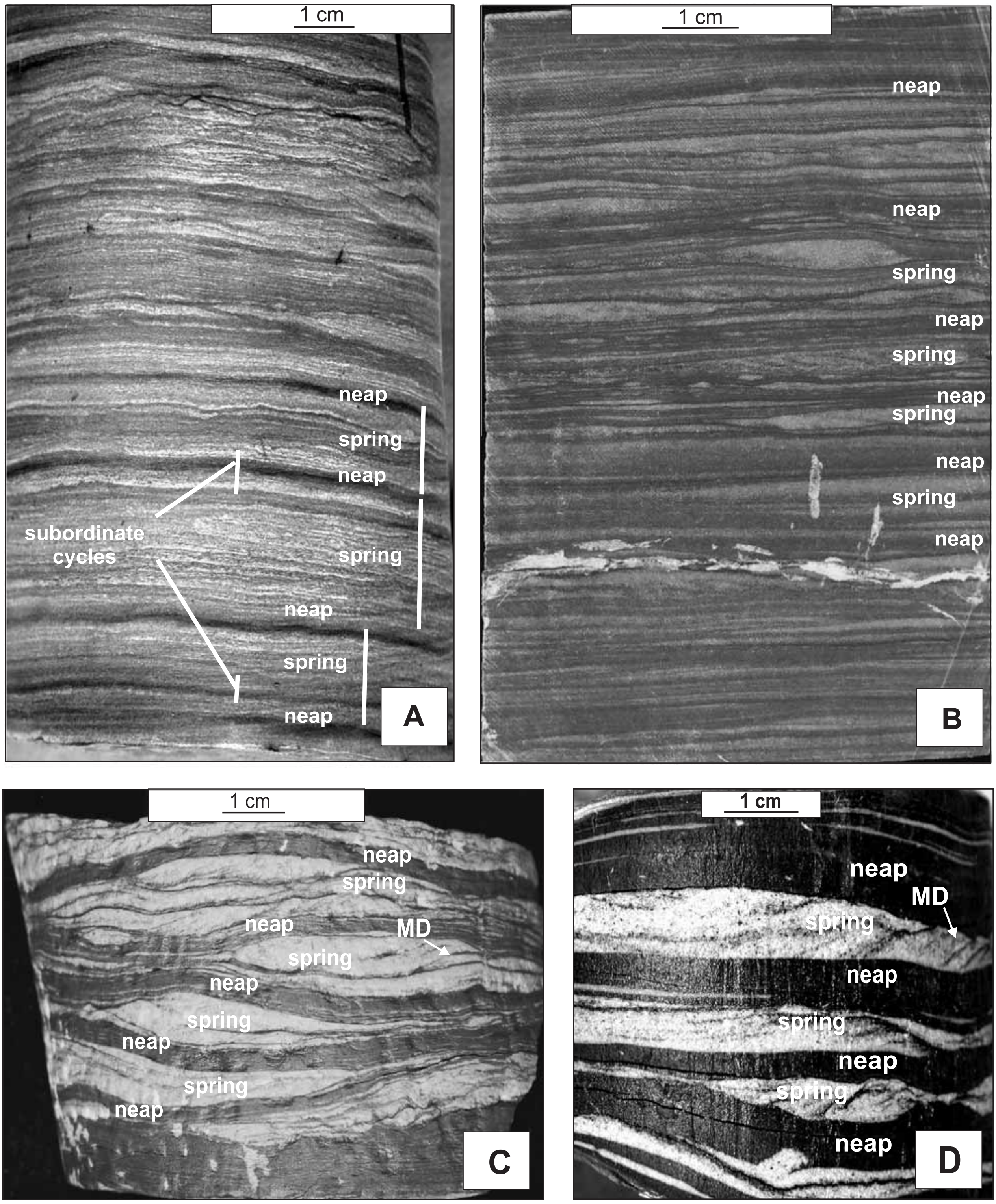

Fig. 46. Tidal rhythmites from mixed tidal flat of the Lublin estuary, late Ediacaran, Łopiennik Formation

A – planar-laminated tidal rhythmites with neap–spring and subordinate cycles, Podedwórze IG 2 borehole, depth 702.6 m; **B** – planar-laminated tidal rhythmites with neap–spring cycles, Łopiennik IG 1 borehole, depth 5474.0 m; **C** – lenticular-laminated tidal rhythmites with neap–spring cycles. Note multiple lamina character of the neap stage of neap–spring cycles and thin mud drapes on cross lamina foreset (MD). Kaplonosy IG 1 borehole, depth 1392.8 m; **D** – ripple-laminated tidal rhythmites with thick neap laminae and mud drapes on cross lamina foreset (MD), Białopole IG 1 borehole, depth 2905.9 m from each other by black mudstone (or claystone beds) 2–3 mm in thickness. There are also mudstone beds 0.5 cm in thick composed of a three to five very thin (<0.5 mm) claystone laminae. A single sequence composed of a tidal bundle and a mudstone or claystone bed represents a vertical record of a neap–spring tidal cycle (Fig. 46A, B). Mudstone beds correspond to neap cycles and represent repeated slack-water periods, when weak tidal currents deposited the mud from suspension (Tessier, Gigot, 1989; Porębski, 1995; Williams, 1998). During spring cycles, very high-velocity tidal currents deposited thin sand-mud couplets. The presence of thin mud layers in spring couplets proves an increased mud content in the middle estuary, as evidenced by the occurrence of the turbidity maximum. A similar cyclicity is observed in modern and ancient tidal successions (e.g. Tessier, Gigot, 1989; Lanier *et al*., 1993; Martino, Sanderson, 1993; Tessier, 1993; Tessier *et al*., 1995; Gradziński, Doktor, 1996; Williams, 1998; Tape *et al*., 2003; Weedon, 2003; Eriksson *et al*., 2006; Rebata-H *et al*., 2006; Hovikoski *et al*., 2008).

Tidal couplets from the inner estuary are characterised by a different structure. Each couplet consists of fine-grained sandstone beds of variable grain size: from very fine to coarser. Mud layers are absent in the couplets, indicating that the upper part of the inner estuary was characterised by a low mud content in contrast to the middle estuary. Tidal currents transported mainly sand material.

Among planar-bedded rhythmites, there are 5–30 cm thick fine-grained sandstone beds. The sandstone shows flaser (Fig. 32A) or lenticular lamination (Fig. 32E). Wavy lamination is much rarer. A similar ripple lamination succession is observed in most beds (Fig. 32E). Lenticular lamination passing upwards into flaser lamination, or ripple cross lamination overlain by flaser lamination occurs at the base of the succession. Thin mud drapes are often observed on the ripples foreset. They are up to 1 mm in thickness (Fig. 33B). Both these ripple sequences prove an increasing tidal range from the neap to spring phase, and a considerable increase in tidal current velocity during spring tides. The presence of both planar-laminated heteroliths and very frequent structures typical of lower flow regime indicates low-energy tidal currents penetrating the gently sloping mixed flat.

A very similar ripple succession is observed in modern tidal rhythmites of a macro-tidal estuary in the Mont-Saint-Michele Bay of Normandy (Larsonneur, 1994; Tesssier *et al*., 1995). A sedimentologically related record of neap- to spring-tide transition can be observed in tidal deposits of the last Holocene transgression in the Yangtze River estuary (Hori *et al*., 2001; Uehara, Saito, 2003).

Significant factor is shape of estuary, which enables deposition of couplets on the mixed flat. Most favourable conditions concentrate in a morphologically restricted area of a funnel-shaped estuary. This specific shape serves to amplify the rank of tides. The late Ediacaran tidal rhythmites of the Lublin estuary can be compared to the modern rhythmites from funnel-shaped estuaries, such as those from the bays of Mont-Saint-Michele in Normandy (Tessier, Gigot, 1989; Tessier, 1993; Tessier *et al*., 1995) or Fundy in the eastern coast of Canada (Dalrymple *et al*., 1991).

Tidal rhythmites of the Lublin estuary are characterised by the presence of relatively numerous erosional reactivation surfaces (Fig. 33C, D). Such surfaces are associated with frequently recurring subordinate currents in the main ebb-flood tidal cycles. As the tide direction changes, subordinate tidal currents partly erode the existing tidal bundles succession and they deposit a new succession upon the created reactivation surface. The new succession shows a slightly different, low-angle inclination. Such a process occurs in case of planar laminated rhythmites. In ripple laminated tidal rhythmites (Fig. 46C, D), reactivation surfaces formed as a result of both partial reworking of mud-draped foreset ripples by subordinate tidal currents and renewed deposition of oppositely oriented ripples composing of sand laminae.

Tidal rhythmites also occur in the upper parts of intertidal zone on the mudflat, especially in the lowermost area adjoining the mixed flat. However, they are much more poorly developed as compared with their equivalents from the mixed flat. The Lublin estuary mudflats are best developed in its southeastern part. The sections of the region reveal a succession composed of a few packages, each composed of mudflat and overlying tidal channel deposits (Figs. 22, 25). This fact suggests the presence of a process of vertical accretion of mudflats as the relative sea level rose (Borrego *et al*., 1995; Cotter, Driese, 1998; Carr *et al*., 2003).

The mudflat facies spectrum is dominated by very fine-grained facies: mostly claystone and mudstone facies *F*. They were deposited from suspension in low-energy environments on the uppermost tidal flats. Only few fine-grained sandstone beds may have been deposited during high tide (Alam, 1995). The similarity among the facies features and lack of landward salt marshes show that the late Ediacaran mudflats resemble their recent equivalents from the southwestern coast of Korea (Frey *et al*., 1989; Alexander *et al*., 1991).

In contrast to the tidal flat channels of the upper estuary, the tidal flat channels from the middle estuary are not evenly distributed within the tidal flat succession. In all the sections under study, tidal channels occur either on the sandflats or on the mudflats. This pattern is particularly well expressed in the Łopiennik IG 1 (Fig. 22), Białopole IG 1 (Fig. 25) and Terebiń IG 1 (Fig. 23) sections located near the presumed mouth of the Lublin estuary.

Middle estuarine channel facies on the mudflat are represented in the upper part of sections by coarse-grained sandstone facies *SCp*, *SCl* and *SCh*. The lack of erosional surfaces inside the channel fill indicates that in most cases the fill was represented only by a single sand sheet element. A simple geometry of the channels' fill resulted from a rapid lateral migration of channels. A well-developed erosional surface lined with rounded mudstone pebbles is observed at the base of most channels. The pebbles were detached from the mudflats that surrounded the channels. A small thickness of the coarse-grained channel fill suggests small depths of the tidal channels.

LOWER ESTUARY

Lower estuary is the outermost zone of the estuarine system dominated by marine processes, including waves and tidal currents. The zone is characterised by landward transport of bedload (Dalrymple *et al*., 1992; Dalrymple, Choi, 2007).

In the southeast of the basin (Łopiennik IG 1, Terebiń IG 5, Horodło 1 and Białopole IG 1), a volcanic series is immediately overlain by black claystones and mudstones with rare interbeds of thick laminated sandstone-mudstone heteroliths. The black claystones contain very abundant Cyanophyta of *Vendotaenia antiqua* forma *quarta* Gnilovskaya. The dark colour of the rocks, numerous small pyrite concretions and very abundant Cyanophyta thallus suggest a low level of oxygenation. The black claystones and mudstones show the highest values of total organic carbon (TOC) of 0.4–0.7% (Pacześna *et al*., 2005) within the estuarine transgressive succession. Enrichment in organic matter is typical of oxygen-depleted zones, especially in initial phases of marine transgressions (Herbin *et al*., 1993; Robison, Engel, 1993). The black claystones and mudstones are directly overlain by low-angle cross-bedded fine- and coarse-grained sandstones with erosional surfaces at the base. These deposits are interbedded with mudstone packages.

The above described features of the black claystones and mudstones that overlie sandstone deposits of high-energy environments indicate that they may have been deposited in a partly or completely protected oxygen-depleted basin of poor water circulation. This could be a central bay of a wave-dominated estuary. The beds of laminated sandstone–mudstone heteroliths can represent fragments of a mixed flat that developed near the central bay. The overlying packages of black claystones and mudstones, encountered in all the sections of the southeastern region (Figs. 22–25), can be the relicts of wave-dominated estuary that developed during the initial stages of marine transgression in the southeastern area of the Lublin–Podlasie basin. It was an ephemeral coastal system. No deposits representing a flood delta and a barrier, typical of wave-dominated estuaries, have been found within the studied intervals.

The proportion of coarse-grained and fine-grained sandstone facies significantly increases upwards in the sections. The deposits show numerous evident indicators of tidal activity, including distinct mud drapes overlying cross-bedding foresets, co-occurring with bimodal cross-bedding (Fig. 29B). The most common facies in sandstones is *SCl*. Facies *SFr* and *SFf* are observed rarely. All the sections included a set of packages of fining-upward sandstones. The lower part of each package is composed of coarse-grained sandstone facies *SCl*. The base is commonly represented by an erosional surface

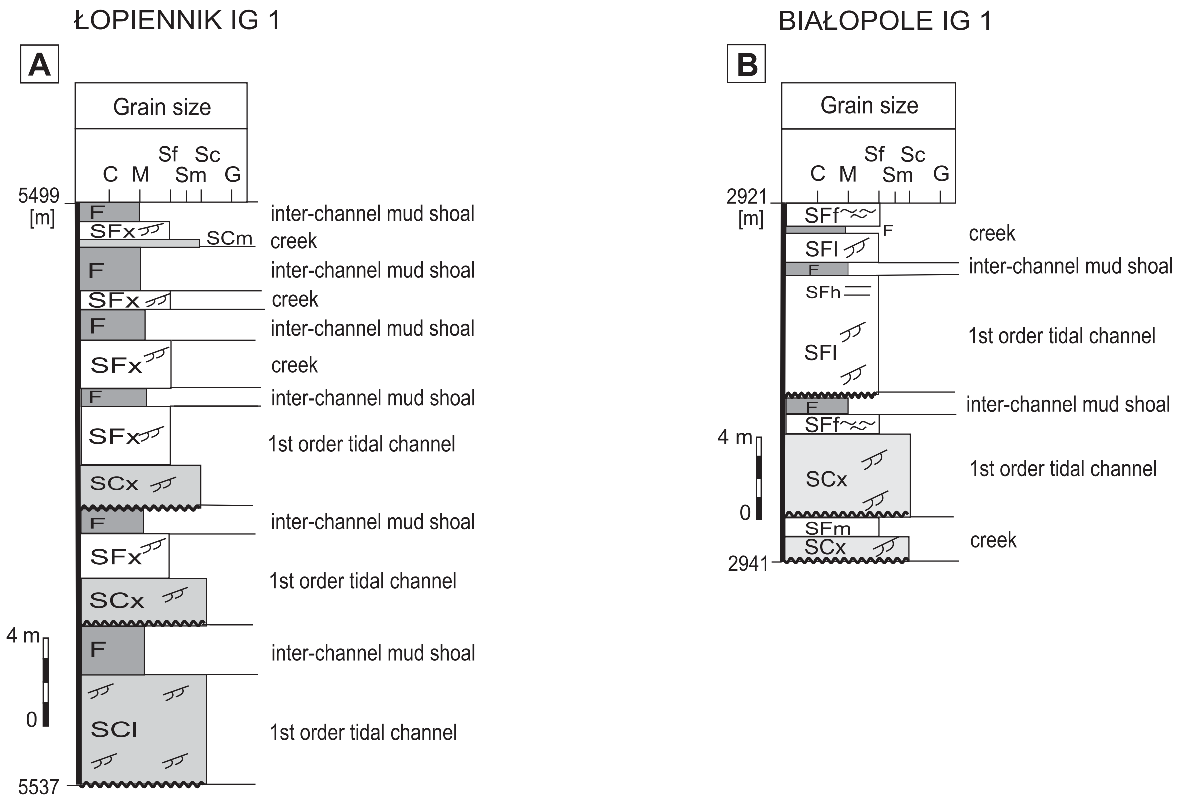

Fig. 47. Facies successions (A) amalgamated tidal channel complex and (B) sand bars in the Lublin lower estuary

For explanations of grain size, sedimentary structures and facies code see Fig. 10

that has up 20–30 cm of relief, lined by mudstone pebbles (Fig. 29C). Sandstone facies *SCl* is overlain by fine-grained sandstone facies *SFr* or *SFf*. In the some packages lower segment is represented by facies *SFl* passing upwards into facies *SFh*. The presence of facies *SFh* and *SFl* suggests upper flow regime conditions due to high velocities of tidal currents. Low energy is suggested by the occurrence of facies *SFr* and *SFf* in the upper parts of the packages.

The presence of tidal indicators, facies architecture and the overlying sandstone packages indicate that mentioned series is a complex of amalgamated tidal channels developed in the subtidal zone. The packages of considerable thickness, sometimes up to 10 m, probably represent 1st order tidal channels. A smaller thickness of packages (2–4 m) can suggest subordinate tidal channels (creeks) (Fig. 47 A, B). In general, each section reveals an upward decrease in the proportion of coarse-grained facies. The channels are filled with fine-grained sandstones suggesting decreasing energy. They fill subtidal channels, which were later converted into channels developed in the intertidal zone, on the sandflat or mixed flat. Most of the packages representing subtidal channels are overlain by 1–3 m thick dark grey mudstone beds. Their origin can be interpreted in two ways. Most likely they represent inter-channel mud shoals deposited by water overflow from the channel onto adjacent areas (Martinius *et al.*, 2001). Another explanation of the origin of the mudstone beds in the upper parts of the subtidal packages is deposition during decreasing flow energy in the tidal channel.

The characteristic elements of the subtidal zone are elongated tidal sand bars. They commonly show some features of oscillatory tidal channels (e.g. Tirsgaard, 1993; Borrego *et al.*, 1995). The frequency of tidal sand bars in the Lublin estuary sections is very low. The lack of well-expressed tidal sand bars in the Lublin estuary sections can be explained by their cannibalization by laterally migrating tidal channels during a rapid rise of relative sea level (e.g. Johnson, 1977; Dalrymple *et al.*, 1990).

EDIACARAN–CAMBRIAN DEPOSITIONAL SEQUENCES

Identification of vertical facies associations stacking patterns, representing system tracts, and key bounding surfaces enabled creating a high-resolution sequence stratigraphic framework reflecting an evolutionary model of the ancient Lublin estuary.

Depositional sequence is defined as a stratigraphic unit composed of relatively conformable succession of genetically related strata and bounded at its top and base by subaerial major discontinuity surfaces representing angular and erosional unconformities and their correlative conformities (e.g. Mitchum *et al.*, 1977; Vail *et al.*, 1984; Posamentier *et al.*, 1988; van Wagoner *et al.*, 1988; Mitchum, van Wagoner, 1991; Wan Yang, Kominz, 2003). First order discontinuity surfaces are subaerial erosional unconformities related to the largest breaks in the facies and stratigraphic spectrum of the geological succession within the basin. The breaks can comprise either part or whole of the geological system, or even chronostratigraphic units of a higher rank. In the Lublin–Podlasie basin, it commonly refers to the late Middle Cambrian and Upper Cambrian in all the regions of the basin, and the early-middle Neoproterozoic in its eastern and southeastern part (Pacześna, Poprawa, 2005a, b). They suggest long periods of subaerial exposure and epigenetic erosion, bound the Ediacaran–Cambrian sections at the top and base and represent the boundaries of 2nd order depositional sequences.

Depositional succession of the Lublin–Podlasie basin commences with siliciclastic sequence A, discontinuously overlying the crystalline basement. Its upper boundary coincides with the lower boundary of sequence B in the Ediacaran–Cambrian succession. The sequences are separated by a pre-rift angular unconformity (Fig. 6). Sequence B includes the late Ediacaran through Middle Cambrian deposits represented by both clastic and volcanigenic rocks. The upper boundary of sequence B is marked by an erosional unconformity spanning the latest Middle Cambrian through early Tremadoc (Pacześna, Poprawa, 2005a, b) (Fig. 4). The basal, pre-rift unconformity in the Lublin–Podlasie basin might be well explained by thermal doming predating the extensional tectonics and volcanism in the region (Fig. 6) (Poprawa, Pacześna, 2002). The lowest part of depositional sequence B is represented by a lowstand systems tract characterised by a low relative sea level that was controlled mainly by local tectonics, such as regional thermal doming and rapid accumulation of alluvial siliciclastic and continental volcanigenic rock, fast enough to compensate the subsidence of extensional grabens. The model of lowstand systems tract is additionally supported by lack of deposition or erosion in regions of the Palaeoproterozoic crystalline basement recorded in the Parczew IG 10 (Fig. 19) and Radzyń IG 1 (Fig. 17) (Pacześna, Poprawa, 2005a, b). Mentioned regions surrounded the spatially restricted zone of deposition of alluvial and volcanigenic rocks. The lowstand deposits are dated at the top at ca. 551±4 Ma (Compston *et al.*, 1995) (comp. Fig. 6).

Within the late Ediacaran and early Cambrian part of sequence B, lower order depositional sequences constitute the transgressive systems tract I (comp. Fig. 6) during the fast relative sea-level rise that gave rise to a marine transgression. Its onset is coincident with the lower boundary of the *Sabellidites-Vendotaenia* Zone (Pacześna, 2008). Because it is impossible to define the time interval during which the volcanigenic series, underlying the transgressive deposits, were subjected to erosion, it can be assumed that the beginning of the transgressive systems tract I was at approximately 551 million years ago.

The Ediacaran–Cambrian succession is terminated by the highstand systems tract I (comp. Fig. 6). It corresponds to a period of relatively intense supply of coarse-grained sand and gravel to nothern part of the Lublin region (Figs. 17–20) and decreasing subsidence (Pacześna, Poprawa, 2005a, b). The upper boundary of the highstand is defined by an erosional surface transgressively modified by a 3rd order depositional sequence boundary. Prior to modification the surface was a lateral equivalent of subaeral erosion, documented east of the Lublin–Podlasie basin in the Ukraine territory (Mens, 1987) The upper boundary of the highstand also approximately coincides with the upper boundary of the *Platysolenites antiquissimus* Zone (Pacześna, Poprawa, 2005a, b). Above, within the upper part of the Cambrian succession, there is another transgressive systems tract II followed by a highstand II. However, these are beyond the scope of this study (comp. Fig. 6).

Nowadays, the commonly accepted rule is that parasequences are the basic elements of the sequence stratigraphic framework. According to the classical definition, parasequences are *relatively conformable successions of genetically related beds or bedsets bounded by marine-flooding surfaces or their correlative surfaces* (van Wagoner *et al.*, 1990).

The extents of the parasequences are limited in occurrence to the Lublin region of the Lublin–Podlasie basin, due to the specific marginal type of the estuarine sedimentary environment (regionally bounded in the north by geomorphological barriers), extent of the estuarine basin, and the rift setting of the late Ediacaran basin.

Parasequences in the estuarine succession have different expressions in individual parts of the estuary. Therefore, the lithologic records of their boundaries, indicating a deepening of the environment, differ from area to area within the estuary (e.g. Holz, 2003). It is due to the very uneven distribution of tidal flat zones in the Lublin estuary at individual stages of its evolution, manifested by the succession of parasequences.

EVOLUTION OF THE LUBLIN ESTUARY

TRANSGRESSIVE SYSTEMS TRACT I – PROGRESSIVE ESTUARY

The parasequences identified within the late Ediacaran transgressive succession are bounded by a flooding surface indicating a deepening of the sedimentary environment. The parasequence boundary of the transgressive phase in the estuary is marked by an abrupt appearance of a deeper tidal flat facies associations that overlie a shallower ones (Figs. 17–20, 22–25, 39) (e.g. Rossetti, 1998; Kim *et al.*, 1999).

Five parasequences, signified PS1–PS5 (from the base to the top: Figs. 17–20, 22–25, 39, 41, 42) have been identified in the transgressive systems tract I that have been distinguished within the sequence B. These parasequences were identified in the Łopiennik IG 1, Białopole IG 1, Terebiń IG 5, Busówno IG 1, Kaplonosy IG 1, Radzyń IG 1, Parczew IG 10, Niwa 1, Święcica 1 and Horodło 1 sections. There is a distinct regional variability in the facies architecture of the two lowermost parasequences PS1 and PS2. Both the parasequences from the sections located in the southeastern part of basin (Łopiennik IG 1, Białopole IG 1, Horodło 1 and Terebiń IG 5), clearly differ from their counterparts from the other regions of the estuary. The lower part of each parasequence is composed of characteristic black claystones and mudstones representing sedimentation under low-oxygen conditions in a partly restricted basin. It was probably a central bay of wave-dominated estuary. Its facies record in the lowermost part of the Lublin estuary succession suggests features of a wave-dominated estuary during the initial phases of transgression. The lack of other elements typical of wave-dominated estuaries, such as flood tidal delta or sand barrier (Figs. 23, 24 and Fig. 48), can be explained by erosional removal during the subsequent events of relative sea-level rise. The occurrence of central bay in the lower estuary, accompanied by the presence of tidal flat in its upper part, suggest a mixed-energy wave-and-tide-dominated type of the Lublin estuary during the initial phases of its evolution. The black claystone and mudstone packages are overlain by relatively thin beds of thinly laminated sandstone-mudstone-claystone heteroliths, representing mixed tidal flat deposits. The above described succession of depositional environments is particularly well expressed in Łopiennik IG 1 (Fig. 22) and Białopole IG 1 (Fig. 25).

In the central part of the middle estuary, near Busówno IG 1, Niwa 1 and Święcica 1, the lower part of parasequence PS1 consists of braided-river and anastomosed river deposits, gradually passing up into estuarine deposits. The estuarine succession of the Busówno IG 1 section is represented by thick (up to 14 m) tidal flat channel fill deposits probably. In Niwa 1 and Święcica 1, there is possibly a complete intertidal succession including sandflat through mixed flat to mudflat deposits. (Figs. 41, 42). The thickness of parasequence PS1 in the central estuary varies from 35 m (Niwa 1) to 50 m (Busówno IG 1 and Święcica 1). The Podedwórze IG 2 drilling was stopped within the upper part of the late Ediacaran succession, thus parasequence PS1 has not been reached in this section (Fig. 18).

In the northwestern and northeastern areas of the Lublin basin, there where the upper estuary developed, parasequence PS1 attains considerable thicknesses ranging from 35 m in Kaplonosy IG 1 (Fig. 20) to 24 m in Parczew IG 10 (Fig. 19). The lower part of parasequence PS1 is composed of a fluvial series represented mainly by braided-river deposits. Alluvial fan deposits compose part of the alluvial succession only in Parczew IG 10 (Fig. 10 B). The upper part of parasequence PS1 is represented by tidal flat deposits (Figs. 17, 19, 20).

The base of parasequence PS1 is an erosional transgressive surface, which clearly marks the onset of sedimentation and development of the Lublin estuary. During the sedimentation of parasequence PS1, a rapid rise of relative sea level gave increase to the development of estuarine sedimentation in a gradually sinking, broad, very shallow incised valley of a river running down SE'wards from the Podlasie region. A shallow fluvial incision in a river valley is observed in the sections situated in the central part of the middle estuary. The incision is recorded there by a well-developed erosional surface in volcanic rocks underlying the braided river deposits, as evidenced by the record in the Busówno IG 1 section (Fig. 39).

Another characteristic feature of the deposition of parasequence PS1 was development of extensive tidal sandflats in the upper areas of the estuary (Figs. 17, 19, 20, 48). The development of sandflats was associated most likely with increased supply of clastics to the upper estuary from strongly eroded source areas situated in the north of the Podlasie basin. The mixed flats and mudflats were very poorly developed at that time and account for approximately 10% to 20% of parasequence PS1. They developed in narrow belts extending along the outer margins of the valley (Fig. 48). The above presented scheme of spatial distribution of sedimentary environments during the initial phases of the Lublin estuary development was related to low influence of tides. The incompleteness of the intertidal zone, especially the poor development of mixed flats with deposition of tidal rhythmites, clearly indicates that the estuary did not exhibit a funnel shape during the initial phase of its development.

The thickness of next parasequence established in the estuary succession, PS2, varies from 8 m in Parczew IG 10 (Fig. 19) to 14 m in Kaplonosy IG 1 (Fig. 20). The parasequence is represented by either an incomplete tidal flat succession without mudflat deposits in Parczew IG 10 (Fig. 19) or a complete tidal flat succession in Kaplonosy IG 1 (Fig. 20). In the central part of the middle estuary, the thickness of parasequence PS2 is greater, ranging between 11 m in Niwa 1 (Fig. 41) to 23 m in Święcica 1 (Fig. 42). The parasequence is composed of mixed flat and mudflat deposits. Parasequence PS2 from the lower estuary shows sandflat, mixed flat and mudflat deposits. The thickness of parasequence varies from 15 m in Białopole IG 1 (Fig. 25) to 6 m in Horodło 1 (Fig. 24).

During the sedimentation of parasequence PS2 there was another phase of expansion of the intertidal zone in the estuary due to the next pulse of relative sea-level rise probably of local tectonic origin (Pacześna, Poprawa, 2005a, b). The zone of relatively extensive sandflats moved to the upper parts of the estuary, whereas in the central and southeastern areas, mixed flats became dominant with the typical sedimentation of tidal rhythmites (Fig. 49). The appearance of thicker packages of tidal rhythmites in the facies spectrum marks the beginning of the funnel-shaped estuary, which determines this sedimentation type in tide-dominated estuaries (e.g. Lanier *et al.*, 1993; Carr *et al.*, 2003). The development of specific hydrodynamic conditions in the estuarine basin was highly influenced by the presumed presence of two crystalline basement elevations in the upper parts of the estuary. The elevations were a natural barrier preventing further development of the estuary into the river valley. Nearby, there was a tidal limit zone (e.g. Figs. 48–50). Simultaneously, the topographic narrowing of the valley caused increase of the tidal rank in the upper part of the Lublin estuary. Mudflats were just a fraction of the intertidal spectrum of the estuary, like during the sedimentation of parasequence PS1.

The thickness and facies map of parasequence PS1 (Fig. 48) shows a characteristic isochore pattern indicating the presence of depressions and elevations of the valley floor at that time. The greatest thicknesses (up to 50 m) are observed near Busówno IG 1, Święcica 1 and Niwa 1 sections. The thickness decreases towards the SE to ca. 4 m near Łopiennik IG 1 and Białopole IG 1 (Fig. 48). Further in this direction, it again increases to 17 m. To the NW of the depocentre located in the centre of the valley, the thickness gradually decreases when approaching the two crystalline basement elevations (Fig. 19). The zonal variability of the thickness pattern may have been caused by tectonic factors. The transgression advanced into the river valley of locally differentiated syndepositional subsidence, stimulating variability of the accumulation space in the estuarine basin (Porębski, 2000; Willis, 2000; Holz, 2003; Takano, Waseda, 2003; Olivero *et al.*, 2008). It should be stressed here that the late Ediacaran transgresssion entered the Lublin basin during the period of fading extensional activity accompanied by a gradual leveling of the rift topography represented by a series of grabens and half-grabens. The development of a vast, NW–SE-elongated rift valley was the ultimate effect of the process. Due to the effect of regionally differentiated subsidence, rivers of different channel patterns could flow in different sections of the valley at the same time as a result of different river's gradient. In the axial part of the valley, where the gradient is least steep, anastomosed rivers flowed. Reduction of the fluvial gradient promoted an evolution in channel styles to anastomosed rivers. In the marginal, north–western and north–eastern areas of the basin, where the gradients were steeper, braided rivers flowed.

The NW–SE direction was typical of the late Ediacaran synrift phase. Tectonic activity during the sedimentation of parasequences PS1 and PS2 may be suggested by tectonic dips (20–25°) observed within two depth intervals in tidal flat deposits of the upper estuary in Kaplonosy IG 1. Although it is impossible to determine confidently the age of the dips, they can suggest high probability of tectonic deformation of the deposits (Pacześna, 2006).

The thickness and facies map of parasequence PS2 presents less accentuated different zones of thickness (Fig. 49). In the areas of Święcica 1 and Busówno IG 1 sections, parasequence PS2 is relatively thick, attaining 23 m. It proves the occurrence of a depositional depocentre in this region, like during the sedimentation of parasequence PS1. The depocentre was related to the fading tectonic activity of the graben extending in the valley basement.

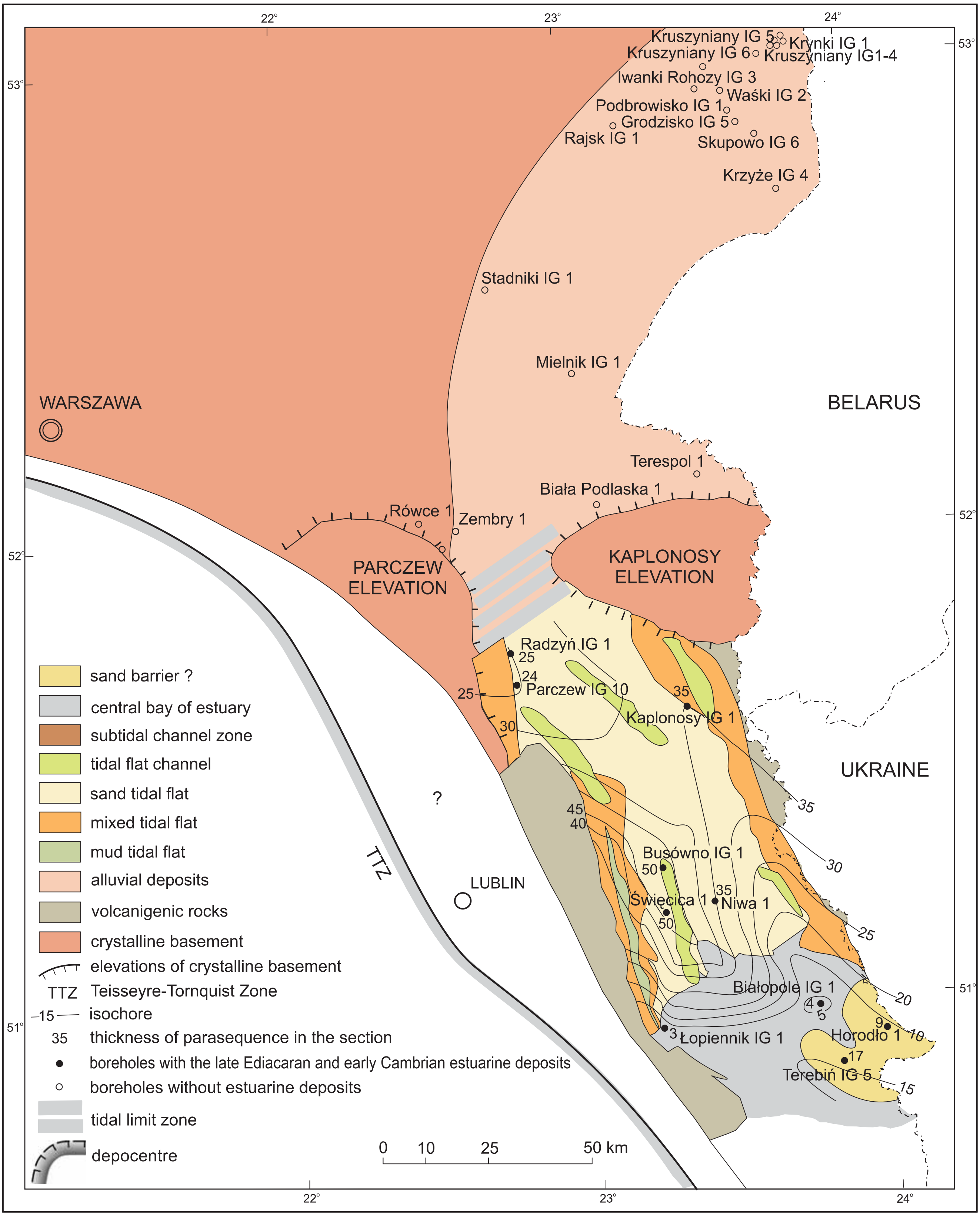

Fig. 48. Combined facies map and isochore map of PS1 parasequence, showing distribution of sedimentary environments in the earliest stage of Lublin estuary development

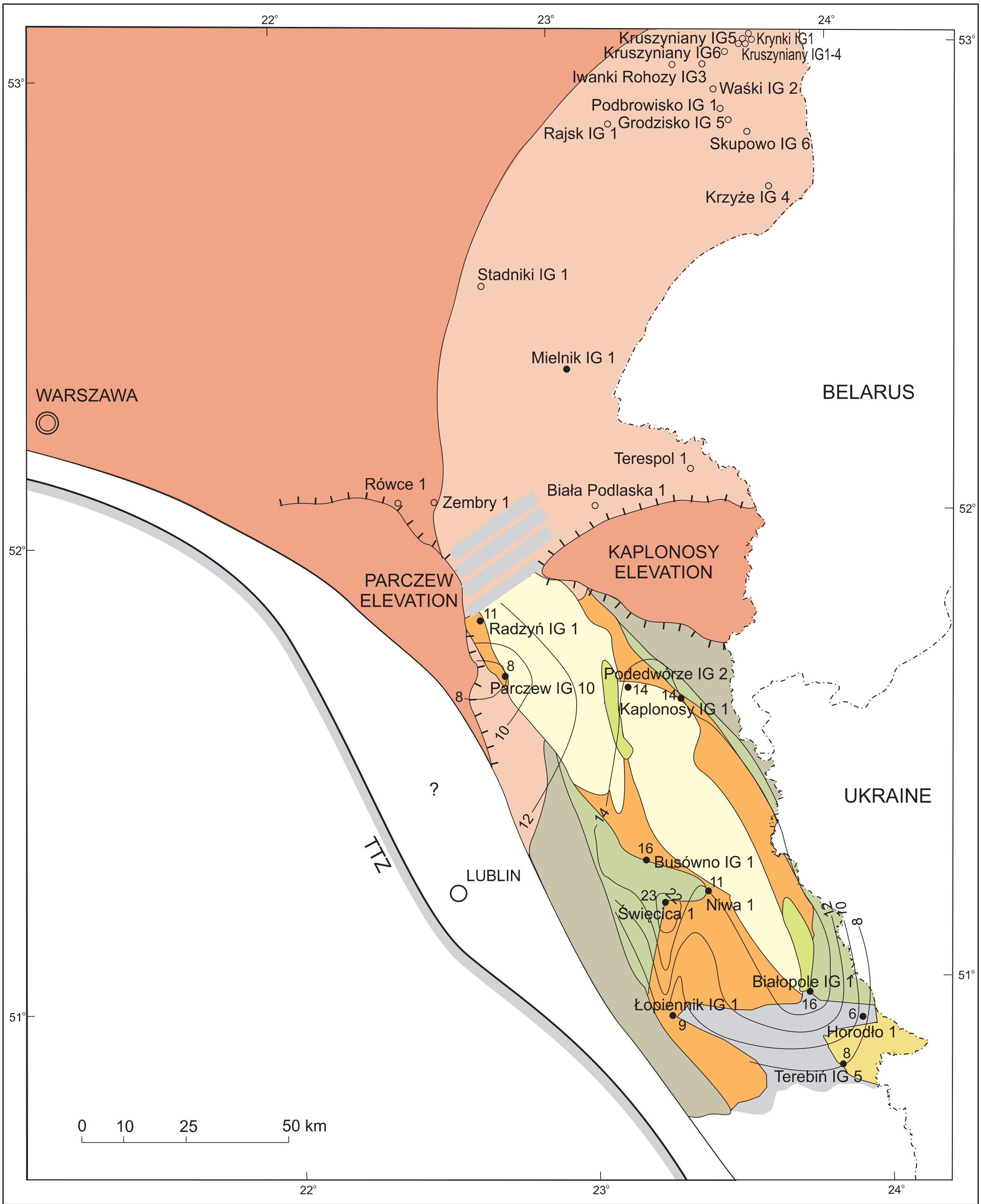

Fig. 49. Combined facies map and isochore map of PS2 parasequence, showing distribution of sedimentary environments in the early stage of Lublin estuary development

For explanations see Fig. 48

The next parasequence in the Lublin estuary succession, PS3, distinguishes among the other late Ediacaran parasequences of the lower estuary region by a very large thickness reaching 106 m (Białopole IG 1) (Figs. 25, 50). The basal part of the parasequence in all the lower estuary sections is composed of a complex of amalgamated subtidal channels of considerable thicknesses ranging up to 22 m (Łopiennik IG 1) (Fig. 22). The complex consists of five cycles representing autocyclicity resulting from migration of subtidal channels (Fig. 47). The lower part of each cycle is composed of fine-grained sandstones showing low-angle cross-bedding and ripple cross lamination. These are subtidal channel deposits. The upper part of the succession consists of mudstones representing inter-channel mud shoal deposits. The subtidal channel complex of parasequence PS3 is overlain by an intertidal zone package with distinctive intervals of thick mixed flat deposits. They are represented by typical tidal rhythmites, characteristic of funnel-shaped estuaries. The package of tidal rhythmites attains a thickness of 65 m in the Łopiennik section (Fig. 22).

In the central part of the middle estuary, parasequence PS3 is much thinner (between 11 m in Busówno IG 1 (Fig. 39) and 34 m in Święcica 1 (Fig. 42). The environmental architecture of parasequence PS3 is simple in this part of the Lublin estuary. Parasequence PS3 is represented by a single package of incompletely developed tidal flat, without the lowermost, sandflat part (Figs. 39, 41, 42).

In the upper estuary, parasequence PS3 is thin. Its thickness ranges from 7 m in Parczew IG 10 to 10 m in Kaplonosy IG 1. Both in the central region of the middle estuary and in the upper estuary, the packages of mixed flat heterolithic deposits are featured by an increased number of thin fine-grained sandstone layers, as compared with the southeastern region of the lower estuary. This phenomenon is associated probably with a longer distance of the lower estuary area from source areas of clastics, situated in the Podlasie sector of the sedimentary basin.

During the deposition of parasequence PS3, when the relative sea-level rose and the transgression advanced into the river valley, the development of mixed tidal flats became more intense (Fig. 50). At the same time, the estuary fully acquired its funnel-shaped geometry. The tidal flat parasequences of the late Ediacaran estuary are of progradational type, as evidenced by the presence of shallower environments of finer-grained sub-associations of the intertidal zone, which appear upwards in the succession (Fig. 51). It means that the uppermost part of the parasequence is composed of mudflat deposits representing the shallowest environment of the intertidal sequence. These are transgressive tidal flats observed in both modern and ancient settings with large supply of detrital material (e.g. Dalrymple *et al.*, 1990; Kim *et al.*, 1999; Mangano, Buatois, 1999). The occurrence of the progradational tidal flat sequence means that it expanded towards the estuary mouth while the transgression advanced.

Tidal flats, prograding simultaneously with developing transgression, occur in estuaries fed by large amounts of clastic material supplied from the land due to fluvial transport to the upper estuary and through transport by tidal currents (Klein, 1985). Since the mixed tidal flats evolved within the fully developed tidal-dominated estuary, the tidal currents were the main factor transporting clastic material up the estuary from the marine basin.

The above discussed processes resulted in the expansion of mixed tidal flats towards the mouth of the Lublin estuary (Fig. 50). These processes were especially intense in the upper estuary. The expansion of mixed tidal flats in both protected and open coasts is associated with increasing subsidence in the basin (e.g. Holz, 2003; Takano, Waseda, 2003). The Lublin estuary sedimentation was controlled by fading tectonic processes in the late stages of synrift phase. Increased subsidence took place in depocentres situated in the valley basement in the rift graben and half-graben zones (Pacześna, 2006).

In the estuary mouth, near Łopiennik IG 1, Białopole IG 1, Horodło 1 and Terebiń IG 5 sections, a wide zone of subtidal channels developed at the same time, as evidenced by the presence of a thick complex of 1st order amalgamated subtidal channels. The complex is the record of migration of channels on the subtidal zone. During the transgression, the lowermost part of the complex with a distinct basal erosional surface is record of a tidal ravinement surface (Allen, Posamentier, 1994), indicating a macrotidal character of the Lublin estuary (e.g. Dalrymple, Zatlin, 1994; Kitazawa, 2007) during the sedimentation of parasequence PS3 (Fig. 52). The surface is especially well pronounced in the lower estuary sections. Because the estuary evolved during the transgression, the upper part of the intertidal zone (including the mudflat) was removed by erosion of tidal channels during the successive events of relative sea-level rise. The process is manifested by the presence of the tidal ravinement surface. This phenomenon can partly explain the lack of supratidal zone within the late Ediacaran succession.

The thickness and facies map of parasequence PS3 shows a gradual increase in the parasequence thickness towards the estuary mouth. This is a distinct record of a tidal prism developing towards the southeast: from the central part of the middle estuary towards its mouth (Fig. 50). This trend suggests that ebb currents were the major factor that contributed to its formation (e.g. Rossetti, 1998). The occurrence of tidal prism is indicative of the complete development of tide-dominated estuary during the sedimentation of parasequence PS3.

Lack of thickness zonation which is characteristic of the underlying parasequences of PS1 and PS2 proves a dying influence of the basement on the valley fill. The small thickness of parasequence PS3 in the upper estuary zone (Fig. 53) is associated with a considerable decrease in the accommodation space near the crystalline basement elevations (Fig. 50).

The next parasequence, PS4, in the lower estuary succession is characterised by variable thicknesses ranging from 8 m in Białopole IG 1 (Fig. 25) to 31 m in Horodło 1 (Fig. 24). It consists of packages representing the completely developed tidal flat. The upper estuary is featured by the occurrence of a parasequence of incomplete architecture. No mudflat deposits are observed in most of the sections located there. The only section with a completely developed succession of tidal flat environments is Kaplonosy IG 1 and Parczew IG 10. In Kaplonosy IG 1 section, parasequence PS4 attains the greatest thickness of 27 m (Fig. 20).

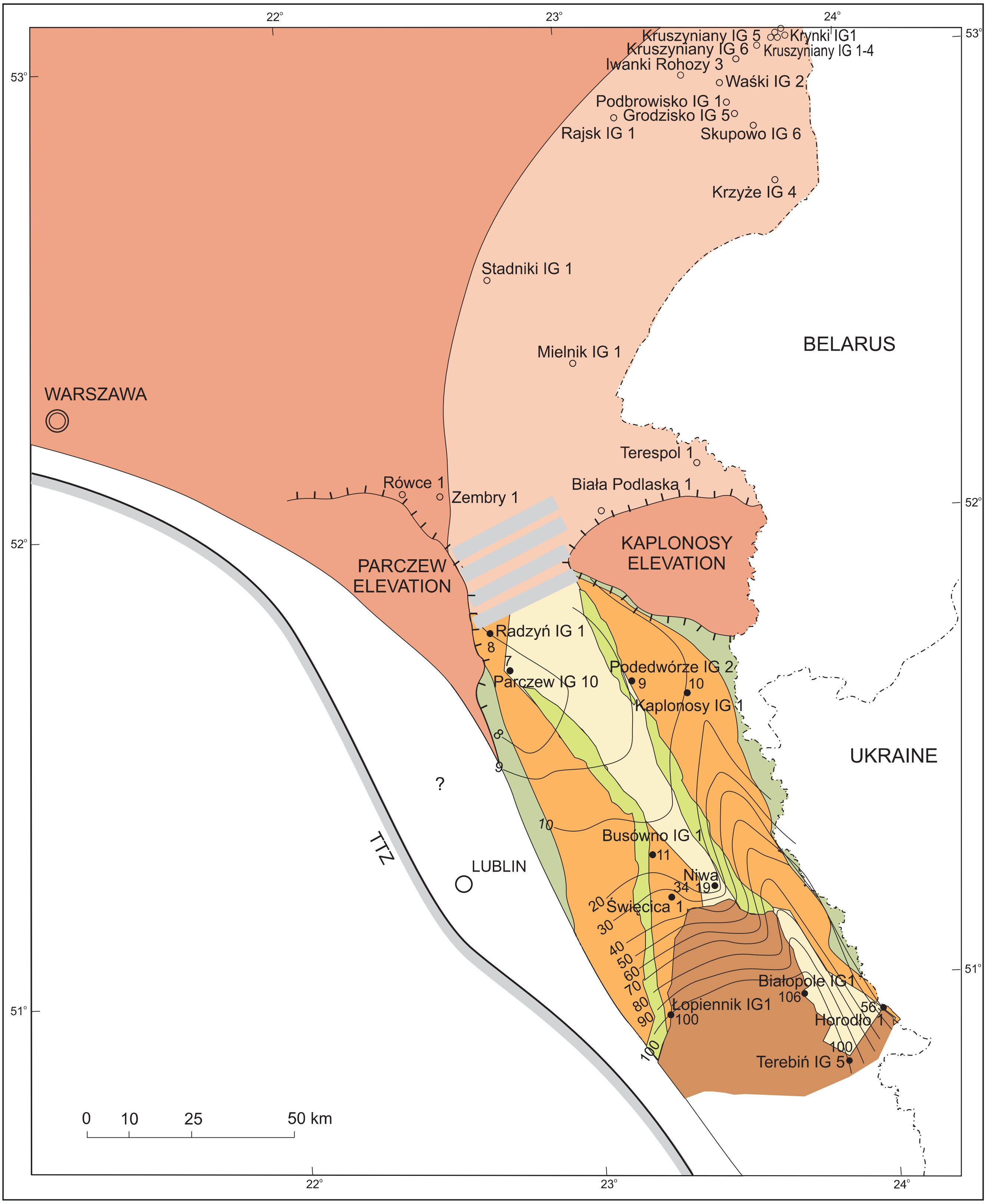

Fig. 50. Combined facies map and isochore map of PS3 parasequence, showing distribution of sedimentary environments in the climax stage of development of the Lublin estuary

For explanations see Fig. 48

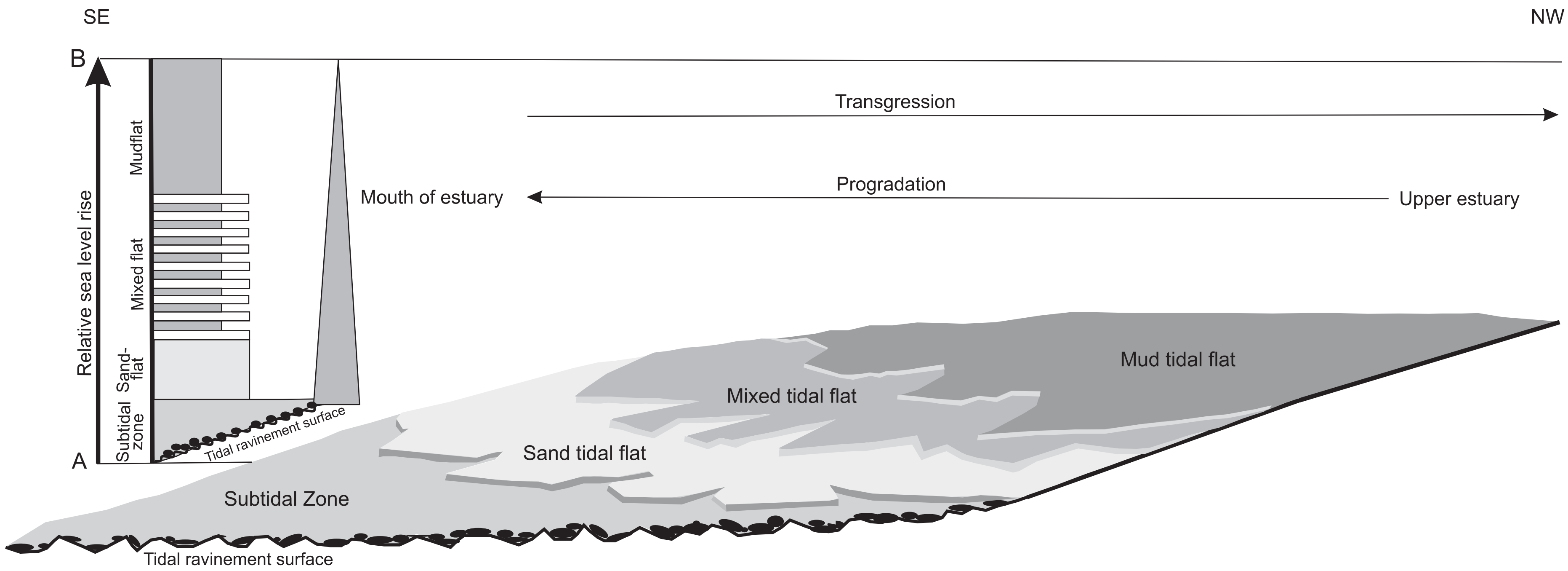

Fig. 51. Schematic cross-section across the Lublin estuary tidal flat, showing the prograding, fining-upward succession of sand, mixed and mud facies associations during relative sea-level rise and increased material supply

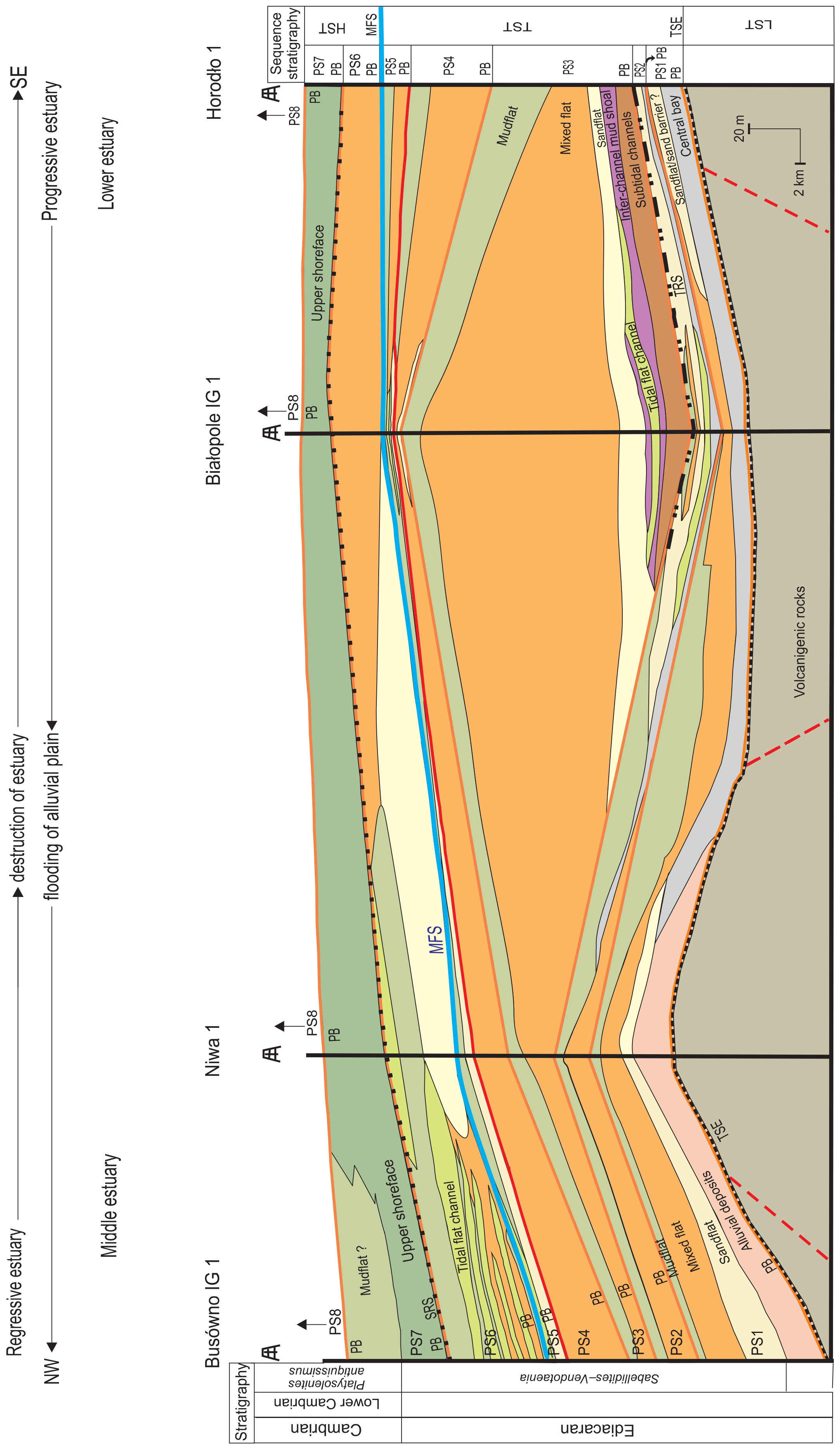

Fig. 52. Facies cross-section along the middle and lower estuary. Note a large thickness of parasequence PS3 caused by increase of subsidence in the rift graben. Datum: maximum flooding surface

Explanations for figures 52 and 53

SEDIMENTARY ENVIRONMENTS

- central bay
- complex of amalgamated subtidal channels
- tidal flat channel
- sand tidal flat
- mixed tidal flat
- mud tidal flat
- alluvial deposits
- proximal upper shoreface
- lower shoreface
- inter-channel mud shoal
- volcanigenic rocks
- crystalline basement
- presumably faults

Parczew IG 10

borehole

SEQUENCE STRATIGRAPHY

LST	lowstand systems tract
TST	transgressive systems tract
HST	highstand systems tract
MFS	maximum flooding surface
PS1	parasequence
PB	parasequence boundary
TSE	transgressive surface of erosion
TRS	tidal ravinement surface
SRS	shoreface ravinement surface

Another tectonically controlled stage of relative sea-level rise initiated a retreat of mixed flats of the lower, middle and upper estuary. The characteristic feature of the estuary evolution is also an asymmetric development of sandflats observed only in the western part of the estuary, along the line of the Podedwórze IG 2 – Busówno IG 1 – Święcica 1 – Łopiennik IG 1 wells (Fig. 54). A significant role in the environmental spectrum of the intertidal zone was played at that time by tidal channels developed on the sandflat and mixed flat (Fig. 54). During the sedimentation of parasequence PS4, mudflats started to develop and culminated during the deposition of the next parasequence PS5.

Parasequence PS5 terminates the transgressive late Ediacaran succession. It is characterised by a highly increased proportion of mudflat packages in the southeastern sections. The greatest thickness (20 m) is observed in the Horodło 1 section (Fig. 24). Alternating packages of tidal flat channels and mudflat mudstones occur in Łopiennik IG 1 (Fig. 22). In the central region of the middle estuary, the environmental architecture of parasequence PS5 was dominated by mudflat and tidal channel deposits observed in the uppermost part of the tidal flat (Fig. 39, 41, 42). In the NE and NW part of the Lublin estuary, parasequence PS5 is characterised by a very small proportion of mixed flat deposits that are observed only in Parczew IG 10 and Radzyń IG 1 as 2 to 4 m thick packages (Figs. 19, 17).

The upper boundary of parasequence PS5 represents the maximum flooding surface I (comp. Fig. 6) in all the sections. In the Lublin estuary, the surface is marked by the maximum extent towards inland of the mudflat deposits, recorded in gamma logs at the depth of peak gamma ray values corresponding to tidal flat mud-clay deposits (Figs. 17–20, 22–25, 39, 41, 42).

The retreat of mixed flats intensified during the sedimentation of parasequence PS5 after a period of sandflat and mixed flat predominance in the Lublin estuary. Mixed flat deposits were encountered only in the sections located along the line of Podedwórze IG 2, Kaplonosy IG 1, Niwa 1 and Białopole IG 1 (Fig. 55). The parasequence PS5 is characterised by the development of the shallowest tidal flat zone represented by mudflat deposits. It was part of the intertidal zone with a well-developed tidal channel network. The intense development of mudflats during the deposition of parasequence PS5 was associated with the maximum landward extent of the estuary. In that case, it was manifested by a total drowning of the river valley and development of the maximum flooding surface (e.g. Cotter, Driese, 1998; Hou *et al.*, 2003). The occurrence of maximum flooding surface coincides with the greatest extent of mudflats in the estuary basin. The progress of the shallowest tidal flat environment is correlated with the end of the transgressive phase of the Lublin estuary.

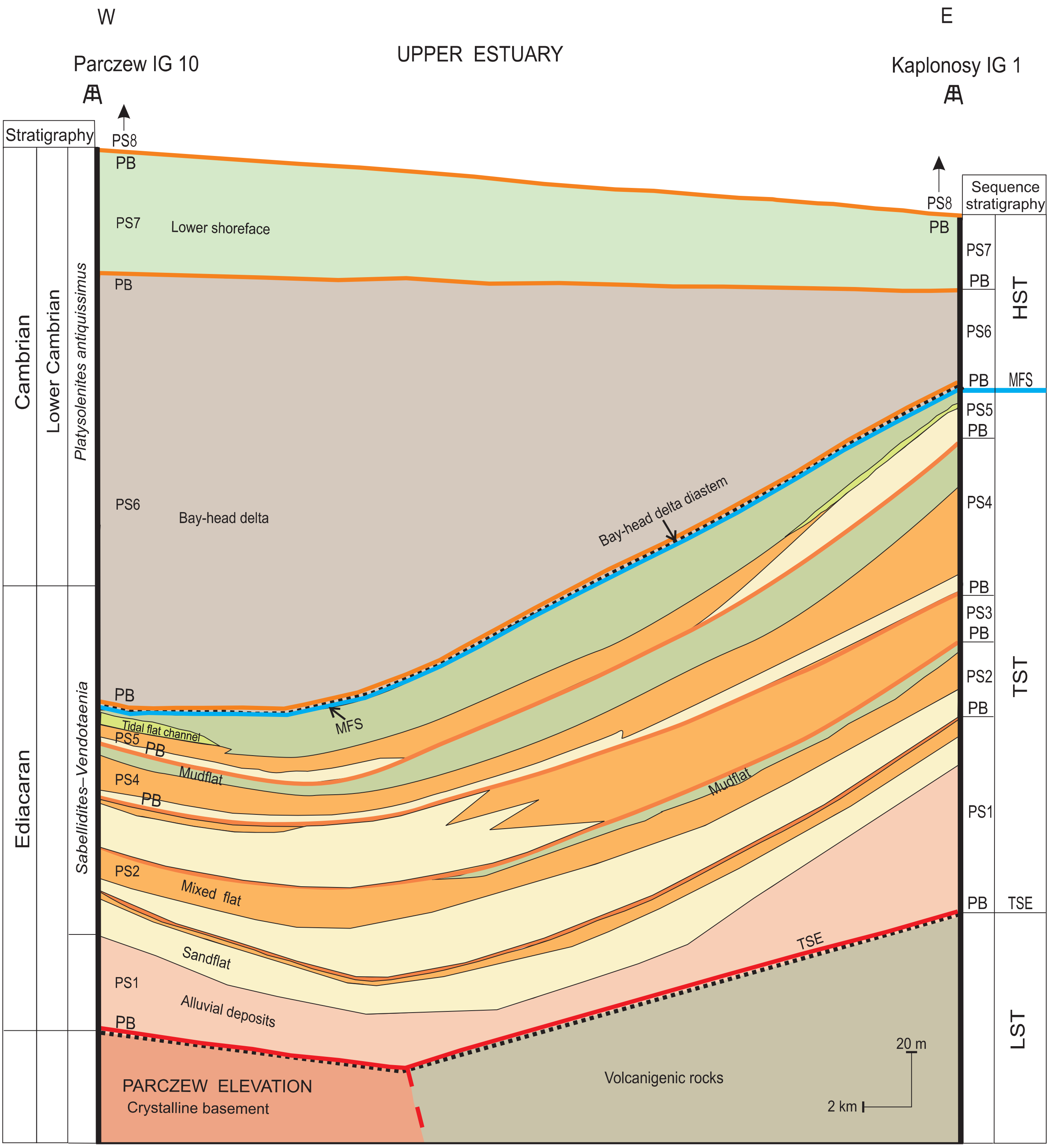

Fig. 53. Facies cross-section across the upper estuary. Note a large thickness of sand tidal flat deposits and parasequence PS6 in the region of Parczew IG 10 section. A large thickness of PS6 parasequence was caused by increase of subsidence in the rift depocentre Parczew–Radzyń during late stage of syn-rift phase. Datum: maximum flooding surface

For explanations see Fig. 52

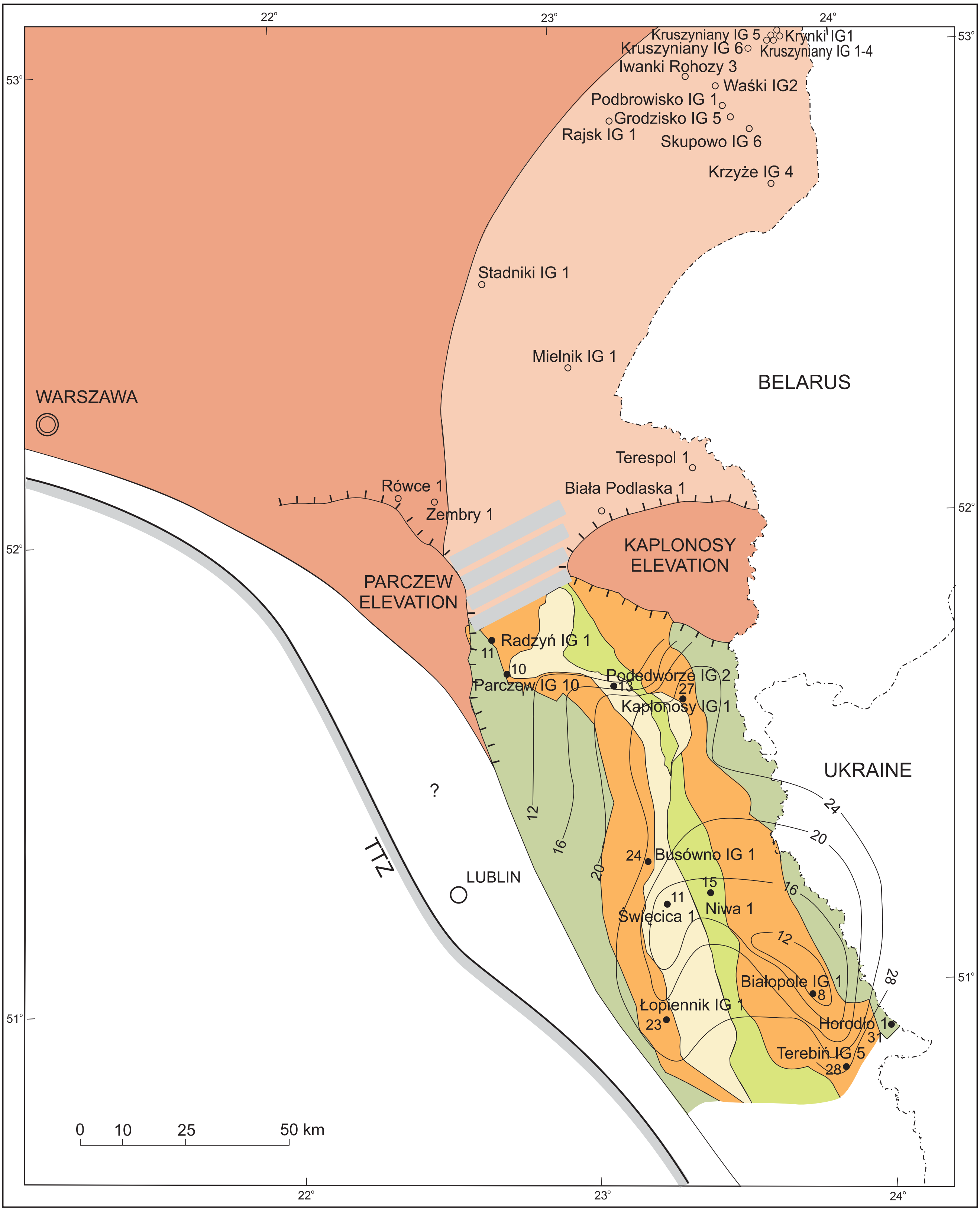

Fig. 54. Late stage of facies development in the estuary-combined facies map and isochore map of PS4 parasequence, showing distribution of sedimentary environments

For explanations see Fig. 48

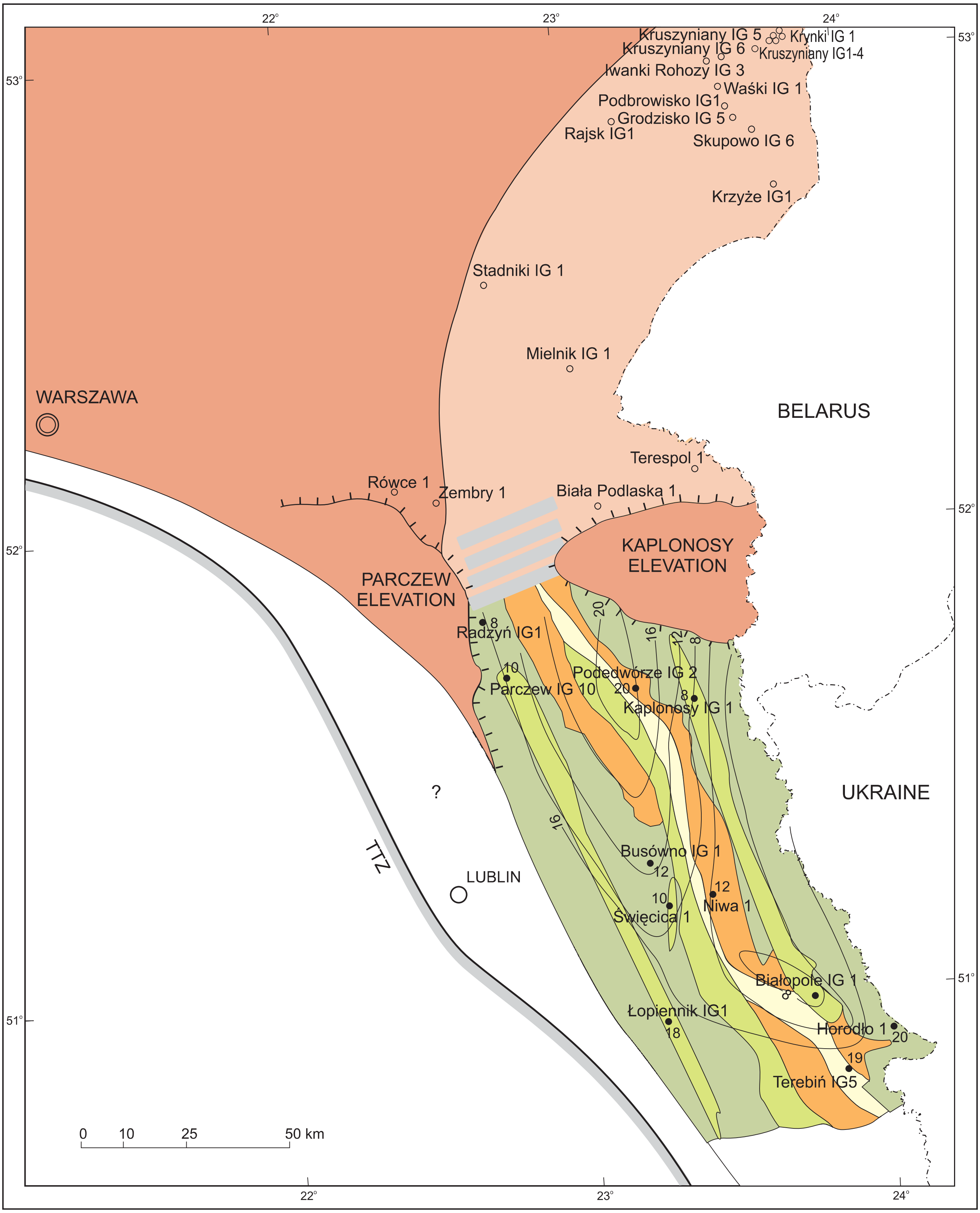

Fig. 55. The beginning of regression of the estuary-combined facies map and isochore map of PS5 parasequence, showing distribution of sedimentary environments

For explanations see Fig. 48

THE HIGHSTAND SYSTEMS TRACT I – REGRESSIVE ESTUARY

In the regressive phase of the estuary evolution, flooding surfaces are defined by the occurrence of nearshore and offshore environments above tidal flat environments in the lower part of the regressive succession. Upper in the succession, the parasequence boundaries are defined by the occurrence of deeper shoreface and offshore facies associations above their shallower equivalents (Figs. 17–20, 22–25, 39).

After the phase of maximum development, a gradual rebuilding of the facies pattern began in the Lublin estuary basin. It is recorded in the four successive parasequences from PS6 to PS9, occurring in the sections above the maximum flooding surface (Figs. 17–20, 22–25, 39, 41, 42). Parasequence PS6 is featured by both a regionally variable thickness and a diverse facies architecture. The greatest thickness of parasequence PS6 is observed in the upper estuary of the Parczew IG 10 section – 79 m (Fig. 19) and in the northern part of central estuary of Busówno IG 1 – 40 m (Fig. 39). Towards the southeast, its thickness gradually decreases from 28 m in Łopiennik IG 1 to 5 m in Horodło 1 (Fig. 56).

In the upper estuary of the Radzyń IG 1 and Parczew IG 10 sections, there is a characteristic series of respectively 4 and 5 sandstone packages in the basal part of parasequence PS6 (Fig. 21). Their lower parts are composed of fine-grained sandstones, whereas the upper parts are represented by either coarse-grained sandstones or conglomerates. Fine-grained sandstones of each package contain low-angle planar cross-bedding or ripple laminations (Fig. 21). The coarse-grained sandstones and conglomerates show low-angle planar cross-bedding. High-angle cross-bedding is much rarer. Immediately above the maximum flooding surface in Podedwórze IG 2, there is a very coarse-grained sandstone layer with low-angle planar cross-bedding, 9 m in thickness. It is directly overlain by a 4 m thick layer of conglomerates. The conglomerate shows very low-angle planar cross-bedding (5–10°). Both the sandstones and conglomerates contain the quartz grains only. This feature proves a very high degree of maturity due to intense reworking of material during transportation. Each mentioned package represents coarsening-upward cycle (Fig. 21). These probably fluvial-sourced deposits formed a prograding bay-head delta of the estuary. The thickness and facies map constructed of parasequence PS6 clearly shows a relatively thick delta in the uppermost part of the estuary. The deposits reveal a distinct trend of decreasing thickness towards the southeast (Fig. 56). This also means that the tidal deposition, so characteristic of the transgressive stage, retreated from the upper estuary. Ripple cross-lamination observed in fine-grained sandstones of the lower parts of the cycles can be the record of tidal currents. No finely laminated sandstone-mudstone-claystone heteroliths, typical of the transgressive stage in the estuary, have been observed in any of the cycles. However, they occur in the lower part of parasequence PS6 in the north of the middle estuary in the Busówno IG 1 section (Fig. 39).

A number of facts support the interpretation that the lowermost part of the regressive estuary succession represents a bay-head delta. These are the overall trend of landward increasing grain size and the gradual basinward increase of the thickness. The presence of bay-head delta is also supported both by proximity to land source areas located in the Podlasie part of the basin and by generally very high maturity of sandstone material, which indicates intense reworking of clastic material during fluvial transport. Another important criterion for the occurrence of bay-head delta deposits in the upper estuary is the presence of an erosional surface developed at the base of a coarse-grained sandstone layer in the lower part of parasequence PS6 (Figs. 18, 19–21). This is most likely an erosional surface termed a bay-head diastem (e.g. Zaitlin *et al.*, 1994; Nichol *et al.* 1994; Roy, 1994). This surface is a record of lateral migration of a river bed in the uppermost part of the bay-head delta. The bay-head diastem migrated basinwards along with the bay-head delta prograding in the same direction. The seaward trend of bay-head diastem migration is well observable in the Podedwórze IG 2, Parczew IG 10, and Kaplonosy IG 1 sections (Figs. 18–20). In Parczew IG 10, located in the north-westernmost area, the bay-head diastem occurs in the latemost Ediacaran just above the maximum flooding surface (Fig. 19). Relatively late, in the lowermost Lower Cambrian Platysolenites antiquissimus Zone, the bay-head diastem occurs in the Kaplonosy IG 1 (Fig. 20) and Podedwórze IG 2 sections (Fig. 18), located in the southeastern ends of the upper estuary.

In most of mixed-energy tide- and wave-dominated estuaries, during the initial stage of their regression, deltaic sedimentation in the upper estuary is accompanied by tidal settings in their middle and lower parts (e.g. Dalrymple *et al.*, 1992; Holz, 2003). A similar distribution pattern of sedimentary environments in the Lublin estuary is indicated by facies architecture variability in the basal parts of parasequence PS6. A lateral equivalent of bay-head delta deposits in the upper estuary sections are tidal flat deposits observed in the northern part of the middle estuary in the Busówno IG 1 (Fig. 39) and, probably, in Niwa 1 and Święcica 1 sections. Due to a very small amount of core material within the interval corresponding to parasequence PS6, the presence of tidal deposits in Niwa 1 (Fig. 41) and Święcica 1 (Fig. 42) cannot be confirmed. Similar relations of simultaneous occurrence of deltaic deposits in the upper estuary and tidal deposits in its lower parts are observed during the initial phases of highstand periods in modern estuaries. A good example is the mixed tide-wave dominated Gironde estuary in southwestern France (Allen, Posamentier, 1994) and the Pleistocene and Holocene estuaries on the eastern shore of Nova Scotia in Australia (Nichol *et al.*, 1994). Studies of ancient estuaries also proved co-occurrence of delta and tidal settings in the initial phases of their regression (e.g. Dalrymple *et al.*, 1992; Holz, 2003; Kitazawa,

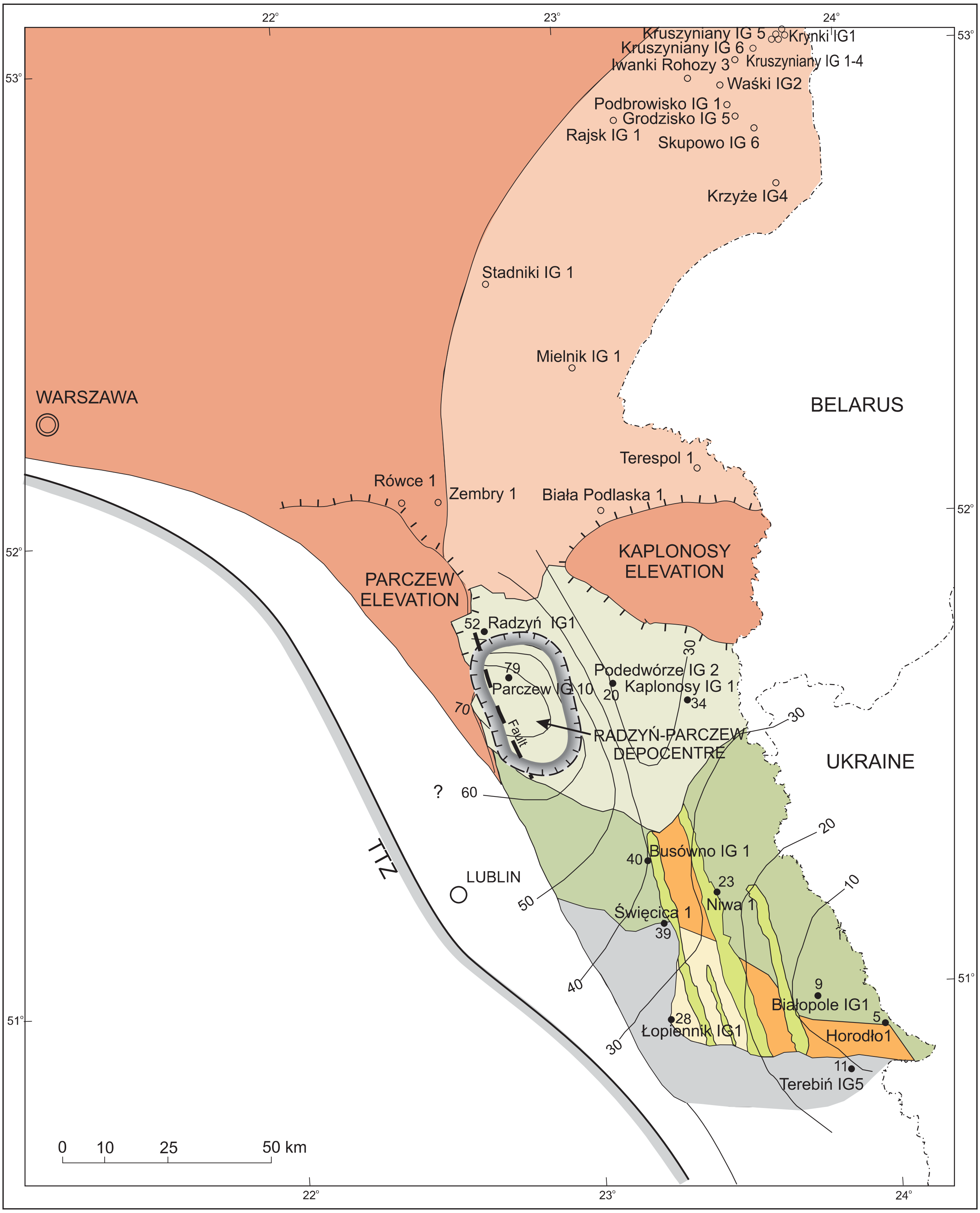

Fig. 56. Combined facies map and isochore map of PS6 parasequence, showing the pattern of sedimentary environments and location of the Radzyń–Parczew depocentre (after Pacześna, 2006, modified) during final stage of development of the Lublin estuary. Note the seaward-prograding bay-head delta in the upper estuary and shoreface in its lower part

For explanations see Fig. 48

2007). Large thicknesses of deltaic deposits in the upper estuary during sedimentation of parasequence PS6 are related to the occurrence of the small Radzyń–Parczew sedimentary depocentre in this region. The depocentre was associated with the final stage of synrift phase (Fig. 56), being a record of increased subsidence rates in a declining rift graben (Pacześna, 2006). The maximum thickness of the depocentre fill was 79 m. It corresponds to the thickness of deltaic deposits observed in the Parczew IG 10 section (Fig. 19). This fact proves that the increased subsidence favoured the thick delta prism development in the upper estuary (Fig. 56).

In the central and mouth parts of the estuary, parasequence PS6 has a different facies architecture. The mouth area is characterised by the well developed mixed flat that is composed of finely laminated sandstone-mudstone-claystone heteroliths highly resembling the late Ediacaran tidal rhythmites of parasequences PS3, PS4 and PS5. The heteroliths occur in the lower parts of parasequence PS6 in Busówno IG 1 (Fig. 39), Łopiennik IG 1 (Fig. 22) and Horodło 1 (Fig. 24). In the Terebiń IG 5 section, the basal part of parasequence PS6 is composed of mudflat deposits (Fig. 23). It indicates strong effects of tides.The mudflat series is overlain by upper shoreface deposits with a well-developed erosional surface at the base of the fine-grained sandstone bed (Figs. 22, 23, 25, 39). Immediately above the erosional surface or within the overlying layer, there are abundant mudstone and claystone clasts. The occurrence of the erosional surface at the contact between estuarine and open-marine basin deposits indicates that this is an erosional shoreface ravinement surface. This erosional surface is typical of wave-dominated estuaries with a well-developed shoreface near the estuary mouth (e.g. Nichol *et al.*, 1994). The shoreface ravinement surface represents the initial position of the shoreline at the onset of the estuary regression. Subsequent episodes of increased supply of clastic material to the estuarine basin caused a shoreline progradation. In Łopiennik IG 1, two shoreface ravinement surfaces occur in two successive parasequences at the base of thick sand ridges (Fig. 22). The shoreline progradation resulted in the development of a broad sand-ridge plain with shoals between sand ridges, where mud deposition took place.

The shoreface ravinement surface in parasequence PS6 is a precursor of total restructuring of the facies pattern in the Lublin estuary that was gradually transformed into a shallow, open-marine coast. When shoreface deposition started, the basinward progradation of the estuary began. The estuarine basin was already partly filled in its upper part with deposits of the basinward prograding bay-head delta (Fig. 56). Simultaneously, destruction of the estuary was enhanced in its mouth area by the development and progradation of sand ridges in the shoreface representing the so-called prograding shoreface (Fig. 56).

During deposition of parasequence PS6, a highstand stage began. The amount of detrital material supplied to the basin exceeded its accommodation space due to deceleration of post-rift regional subsidence rate, which resulted from a decline of active rift grabens and half-grabens (Pacześna, Poprawa, 2005a, b; Pacześna, 2006).

The boundaries of the next parasequences (PS7, PS8 and PS9) were determined in the borehole sections on the basis of variability in the ichnological and sedimentological records suggesting a deepening of the marine basin on a regional scale. It is manifested by the appearance of both more deep-marine ichnologic-sedimentary associations above shallow-marine trace fossil assemblages and related sedimentary structures (Pacześna, 2001). In contrast to the late Ediacaran tidal equivalents, they represent parasequences indicating an upward increase in grain size, thus corresponding to the classical definition of parasequence (e.g. van Wagoner *et al.*, 1990; Emery, Myers, 1996; Porębski, 1996; Coe *et al.*, 2003; Pieńkowski, 2004).

Compared with the underlying parasequence PS6, parasequence PS7 has relatively unified facies architecture throughout the whole Lublin basin. In the upper estuary, it is composed of upper shoreface and lower shoreface deposits (Figs. 17–20). In the northern part of middle estuary, the interval corresponding to parasequence PS7 was either very poorly cored (Busówno IG 1) or completely coreless (Niwa 1 and Święcica 1). the lithology of uncored intervals was reconstructed in these sections from the analysis of wireline logs, predominantly gamma ray. They show that the interval covering parasequence PS7 is represented by alternating thick mudstone and thin sandstone packages in the Busówno IG 1 (Fig. 39) and Święcica 1 (Fig. 42) boreholes. The gamma ray log in the Niwa 1 section shows a blocky shape with a distinct progradational trend in the lower and uppermost parts of parasequence PS7 (Fig. 41). Direct interpretation of the depositional environment is impossible in these intervals of the succession due to the lack of cores. The lithology and lithologic succession can be ambiguously interpreted from the wireline logs. They roughly suggest the presence of upper shoreface, and mud flat deposits (?) in Busówno IG 1 (Fig. 39), upper shoreface deposits in Niwa 1 (Fig. 41) and alternating upper shoreface and lower shoreface deposits in Święcica 1 (Fig. 42). However, univocal identification of the nearshore environmental zones is impossible using only the analysis of wireline logs.

Parasequence PS7 reveals a facies record of the final evolutionary stage of the Lublin estuary. It is present only in the southeastern region where thin packages of mudflat deposits are observed in the upper part of parasequence PS7 in the Łopiennik IG 1 section (Fig. 22) and in the central part of the basin in Busówno IG 1 section (Fig. 39). At the end of deposition of parasequence PS7, the Lublin estuary basin was completely filled. It is proved by a uniform sedimentary record of a shallow, open-marine basin in the all studied sections.

The next parasequence, PS8, shows a very remarkable trend of upward-shallowing sedimentary environments, accompanied by a grain size increasing. In well cored Białopole IG 1 section, parasequence PS8 is composed of a 15 m thick mudstone packages interbedded with thin fine-grained sandstone layers deposited in the upper offshore and lower shoreface. In the sections where parasequence PS8 is poorly cored Horodło 1 (Fig. 24), Terebiń IG 5 (Fig. 23) and Łopiennik

IG 1 sections (Fig. 22) in the southeastern area of the basin, the interpretation of sedimentary environments was made by a correlation with the marker section of Białopole IG 1. A transition from upper offshore to lower shoreface deposits is commonly observed in most of these sections.

The extent of parasequence PS9 reaches beyond the geologic time interval discussed in this report, stretching out into the Lower Cambrian *Schmidtiellus mickwitzi* Zone. In most of the sections, this interval is represented by upper offshore deposits. Another environmental architecture of parasequence PS9 is observed only in the Busówno IG 1 section where lower shoreface is present (Fig. 39). No set of typical indicators of oscillatory tidal currents, such as reactivation surfaces, mud drapes, bimodal cross-bedding and cyclic tidal rhythmites, have been found in any of the analysed sections. However, a record of wave action is commonly observed in the sections, including signs of storm activity.

The relatively uniform facies architecture of parasequences PS7 and PS8 indicates an initial aggradation of the Lublin basin shoreline. The succeeding parasequences, PS8–PS9, show a general trend of shallowing sedimentary environments towards their tops, accompanied by an increase in the content of coarse-grained deposits. The above features prove a progradational character of these parasequences. It supports the assumption that the interval of parasequences PS7–PS9 can be interpreted as a highstand. It developed as a result of local increase in clastic material supply to the basin without support of both tectonic and eustatic factors. The amount of material supplied to the basin was greater than the increase in its accommodation space due to deceleration of post-rift regional subsidence, associated with the termination of tectonic activity of rift grabens and half-grabens. The relative sea-level rise became much slower at that time (Pacześna, Poprawa, 2005a, b; Pacześna, 2006).

CONCLUSIONS

1. After the cessation of volcanic rift processes, sedimentary depocentres developed in the northern and central parts of the Lublin–Podlasie basin. They were being filled with alluvial clastics. At the same time, a relatively small initial depocentre of Białopole–Terebiń developed in the southeastern part of the basin. Subsequently, a more extensive depocentre of Kaplonosy–Terebiń formed towards the north–west. Due to both a gradual relative sea-level rise stimulated by tectonic factors and resulting marine transgression, the depocentre became successively filled with deposits of two-fold origin. Continental fluvial material was supplied to the basin from the north. In the south, material transported by sea currents predominated.

2. The late Ediacaran Lublin–Podlasie sedimentary basin was subdivided into four facies regions, which differ between one another in the proportions between coarse-grained (coarse-grained sandstones, gravelstones, and conglomerates) and fine-grained lithofacies (represented by claystones, mudstones and fine-grained sandstones). The northeastern and eastern part of the Podlasie Depression and the northern areas of the Lublin slope of the East European Craton were characterised by sedimentation of continental siliciclastics, which lasted in the area until the end of Ediacaran and continued as late as earliest Cambrian times. The central region spans the northern part of the Lublin slope of the East European Craton. The post-volcanic Ediacaran silicislastic section consists of two sedimentary stages here. The lower one is represented by sedimentation of coarse- to fine-clastic continental deposits. The upper stage is characterised by the occurrence of tidal deposition. The central region covers the southeastern part of the Lublin–Podlasie basin. The late Ediacaran and earliest Lower Cambrian siliciclastic succession in this area is wholly represented by tidal deposits.

3. Identification of facies associations and their succession in the sections located in the above mentioned facies regions enabled defining alluvial, estuarine and open coast depositional systems.

4. The alluvial system developed during the late stages of synrift phase in successively developing sedimentary depocentres.

5. The two-member alluvial depositional system occurs in the lower part of the sections immediately above either the volcanigenic rocks or the crystalline basement. Its lower member is composed of poorly rounded and sorted very coarse-grained alluvial fan sediments. They were a depositional element of transversal drainage paths of rift valleys. Its deposition in the northeastern area of the basin was associated with synsedimentary fault activity in the Iwanki Rohozy–Krzyże sedimentary depocentre. Another region of alluvial fan deposition was the area situated immediately close to the slopes of the crystalline basement near Parczew IG 10.

6. Braided river deposits constituted the upper member of the alluvial depositional system in the marginal, western and northern regions of the basin. These were the areas of fluvial deposition in distal parts of alluvial fans. Large braided rivers flowing down transverse to the axis of the sedimentary basin.

7. Braided channels bear features of typical distal to middle reaches sand-bed braided channels. The distinct predominance of very coarse-grained sandstone facies confirms braided, sand-bed nature of the rivers. Another important indicator of its braided character is the common presence of an index sedimentary structure low-angle planar cross bedding characteristic of transverse bars. The presence of massive, coarse-grained sandstone facies representing supercritical flow conditions suggests deposition in high-energy, shallow braided channels with strong aggradation of sediments.

8. The group of coarsest-grained sandstone and conglomerate sections located near the source areas includes Krzyże IG 4, Kaplonosy IG 1 oraz Parczew IG 10 sections. Detrital material was supplied to the drainage area of alluvial fans and

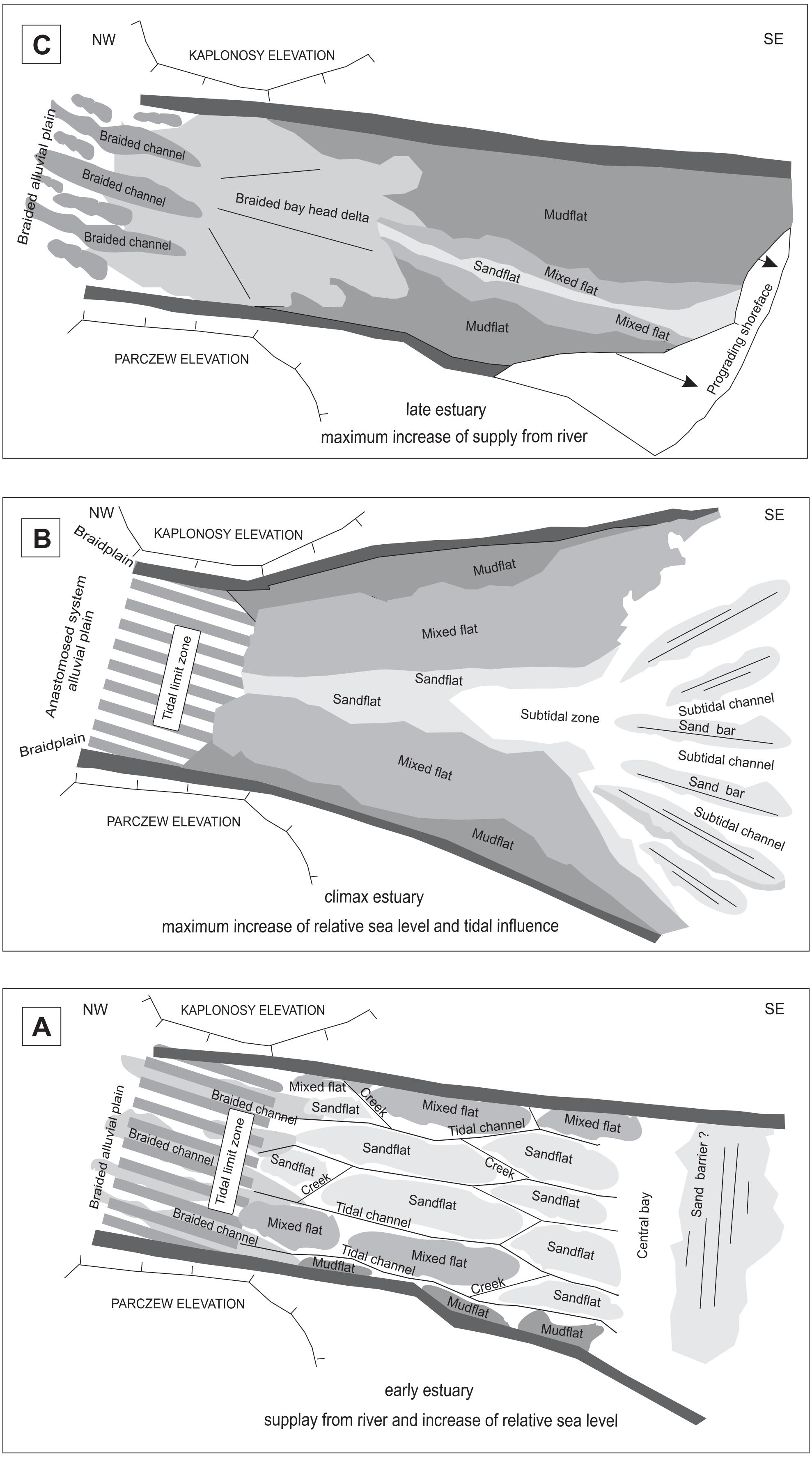

Fig. 57. Three main phases of the Lublin estuary development

A – PS1 and PS2 parasequences mark the birth of mixed energy tidal-wave-dominated estuary; **B** – PS3, PS4 and PS5 parasequences represent the time of maximum development of macrotidal funnel-shaped estuary; **C** – PS6 and PS7 parasequences record the decline of mixed energy tide- and wave-dominated estuary due to covering with bay-head delta in the upper estuary and prograding shoreface in the lower estuary braided rivers flowing in this region from strongly weathering the Palaeoproterozoic and Mesoproterozoic magmatic and metamorphic rocks of elevated areas. These were the Mazury–Belarus Elevation for the Krzyże IG 4 and the Parczew Elevation for the Parczew IG 10 section. The presence of coarse-grained fraction in the fluvial section of Kaplonosy IG 1 supports the suggestion of the occurrence of a crystalline basement elevation in the immediate neighbourhood of the section, called herein the Kaplonosy Elevation. More distal braided channels are observed in Busówno IG 1.

9. Levelling of the rift basin topography and increase in regional subsidence rate of the depocentres in the southeastern area at the end of the alluvial basin evolution resulted in the development of axial drainage of the anastomosed system rivers. The rivers flowed down along the basin axis from north to south.

10. The presence of anastomosed river deposits is documented in the Niwa 1, Święcica 1, Radzyń IG 1 and Mielnik IG 1 sections. They are featured by the distinct predominance of fine-grained sandstone channel facies. Another feature is presence of single-storey channels and excellently developed floodplains. Floodplain deposits of the anastomosed system are characterised by a large thickness as compared with a small thickness of channel deposits. Vertical sediment accretion is a depositional process typical of anastomosed channels. It is evidenced by a very high frequency of massive and low-angle planar cross-bedded sandstone facies. Vertical accretion is also suggested by the lack of fining upwards in grain size within the channel deposits.

11. Changes in the hydrological regime of braided rivers from ephemeral to perennial, development of floodplains of anastomosed river with well-developed crevasse splay deposition and the change in colour of the accumulated deposits probably represent a record of late Ediacaran climatic changes in Baltica from dry and arid to more humid and moderate conditions. Climatic changes were probably the result of the Baltica drift about 560–545 Ma from moderate southern latitudes to palaeoequator zone. Climate humidity increased as the late Ediacaran transgression advanced onto Baltica.

12. At the decline of Ediacaran the marine transgression progressed in the southeastern and central area of the basin. In the northeast, continental sedimentary conditions persisted until earliest Lower Cambrian times.

13. Remarkable variability in the facies spectrum above the late Ediacaran alluvial siliciclastic succession, suggesting a three-zone lateral distribution of facies associations, is one of the main arguments for its estuarine origin. In the northern part of the basin (Radzyń IG 1, Parczew IG 10 and Kaplonosy IG 1), there is a several-metres thick series directly overlying the alluvial succession. The succession of facies associations indicates transition zone, where tidal and river processes interact. The frequency of tidal indicators in very thin packages of mixed tidal flat deposits is very low. They are represented by mud drapes and cyclic tidal rhythmites of limited extent, suggesting the presence of tidal bundles. There was probably a braided tidal-channel network in this zone. It reflected the braided nature of the river system supplying clastic material to the northern areas of the basin, proving that the zone was still under strong influence of fluvial conditions and much more poorly marked tides in the upper, inner estuary.

14. The region of the Busówno IG 1, Niwa 1 and Święcica 1 sections, situated in the central part of the Lublin basin, is characterized by the dominance of finest-grained sediments. This fact is very likely to suggest that the bedload convergence zone was located in this part of the basin. Another specific feature is a south-easterly directed considerable increase in the frequency of tidal indicators, in particular mud drapes occurring in estuaries below the elevation of turbidity maximum zone. The predominance of mixed tidal flat deposits with tidal rhythmites suggests its complete development. Thick sandflat packages showing a sedimentary record of upper flow regime conditions are observed in the southeastern part of the basin near the estuary mouth (recorded in Łopiennik IG 1, Terebiń IG 5, Horodło 1 and Białopole IG 1). It indicates an area of the highest power of the tidal currents, probably indicating location of the maximum tidal range zone in the basin. The presence of the three specific zones: bedload convergence, turbidity maximum and maximum tidal range supports the presumption that the area extending between the lines of Busówno IG 1-Niwa 1-Święcica 1 and Łopiennik IG 1 – Terebiń IG 5 – Horodło 1 – Białopole IG 1 represented the middle estuary zone.

15. Depositional elements of the subtidal zone are observed in the southeastern part of the basin. These are represented by well-developed complex of amalgamated subtidal channels and very poorly developed tidal sand bars (?) with a record of bimodal cross-bedding. They prove the presence of the lower estuary zone which developed within the range of strong marine influences.

16. The late Ediacaran siliciclastic succession of the Lublin basin represents a transgressive stage of the Lublin estuary evolution. It includes five successive episodes of relative sea-level rise recorded in parasequences constituting the transgressive systems tract I. Development of the transgressive systems tract I was controlled by both regional tectonic processes and a considerable increase in material supply to the basin. The evolution of the estuary resulted from a gradual drowning of the NW–SE-trending rift valley that developed during the final stage of the synrift phase of the late Neoproterozoic rift in the western margin of Baltica.

17. The first phase of estuarine sedimentation is recorded in parasequences PS1. During the earliest stages of its evolution, the Lublin estuary was a mixed energy wave-and tide-dominated estuary (Fig. 57A). The mixed character is proved by the following facts: 1) presence of central bay deposits in the lower estuary, 2) simultaneous occurrence of tidal deposits represented by sandflats, and mixed flats with tidal rhythmites, 3) presence of the tidal limit zone in the upper estuary.

18. In the next stage of the transgressive phase, corresponding to the deposition of parasequence PS2, influence of tidal sedimentation significantly increased. It is proved by the occurrence of well-developed mixed flats with tidal rhythmites. Their increasing content in the facies spectrum of the estuary was determined by its gradually forming funnel-shaped geometry associated with the increasing rank of tides in the hypersynchronous tide-dominated estuary.

19. The maximum of mixed flat development, containing typical tidal rhythmites with bundles showing neap-spring tidal cycle, occurred during the deposition of parasequence PS3. It was related to the maximum development of funnel-shaped geometry of the estuary (Fig. 57B). The greatest development of tidal rhythmites in the transgressive succession, presence of tidal ravinement surface and increasing thickness of the tidal prism towards the estuary mouth suggest a macrotidal character of the Lublin estuary during the deposition of parasequence PS3.

20. A regress of mixed flats in the estuary occurred during the deposition of parasequences PS4 and PS5. Development of the maximum flooding surface at the top of parasequence PS5 coincided with the maximum landward extension of mudflats and the ultimate drowning of the river valley with braidplain.

21. Regression of the Lublin estuary started at the Ediacaran/Cambrian transition. The process is recorded in parasequences PS6 and PS7. Regional variability of the facies spectrum of parasequence PS6 is recorded by the simultaneous occurrence of relatively thick upper estuary bay-head delta deposits, middle and lower estuary tidal flat deposits and shoreface zone near the estuary mouth. This fact suggests a change from a macrotidal estuary during the peak stage of the transgression development to a mixed energy wave- and tide-dominated estuary in the initial stage of its regression (Fig. 57C). The decelerating relative sea-level rise and related shoreline progradation marks the beginning of a highstand. Due to increased material supply by a braided river system flowing down from the Podlasie region to the Lublin one, the funnel-shaped topography of the estuary was gradually destroyed.

22. In the *Platysolenites antiquissimus* Chron, the estuary underwent transformation into an open coast highly affected by wave action and, maybe, with a slight influence of tides. At the end of deposition of parasequence PS6 and during the beginning of deposition of parasequence PS7, the Lublin estuary was filled with sediment.

23. The next parasequences derived from regressive phase, PS8– PS9, bear the features of classical parasequences by an increasing grain size and a remarkable upward shallowing of the depositional environment. The sedimentation of open coast deposits in the shoreface to offshore zone occurred, with a record of storm conditions. The sedimentary conditions are proved by the occurrence of a very abundant and ichnotaxonomically diversified assemblage of trace fossils. The early Cambrian trace fossils differ from their late Ediacaran equivalents in more complicated morphology of burrows. Especially they show much larger burrow diameters indicating a well-oxygenated, but low-energy and poorly variable living environment of the tracemakers.

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