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ORIGINAL PAPER

# Fluvial responses to the Weichselian ice sheet advances and retreats: implications for understanding river paleohydrology and pattern changes in Central Poland

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**Abstract**   The evolution of the fluvial systems during the Weichselian Pleniglacial in the Toruń Basin (Central Poland) was investigated through sedimentological investigation and paleohydraulic analysis. Within the basin, three fluvial cycles deposited after successive phases of the ice advance which took place 50, 28 and 20 ka ago. Successions of four fluvial lithotypes characterize each fluvial formation, that are related to the paleoenvironmental changes (e.g., climate instability and changes in the river regime) which affected the channel hydraulics and morphology. The successions comprise river-style metamorphosis between high-energy sand-bed meandering rivers (lithotype M1), high-energy sand-bed braided rivers (lithotype B1), and medium-energy sand-bed braided rivers with either unit bars (lithotype B2) or compound bars (lithotype B3) reflects the maturity stage of sand-bed-braided river evolution in the basin. The assessment of the fluvial sedimentary environments enabled the construction of a quantitative model of the changes in the river channel pattern in relation to the climate oscillation. Both the paleohydrological controls and their sedimentary consequences are discussed in the article. Lithotypes M1 and B1 represent riverbed modeled under supercritical flow condition. Deposition of lithotype B2 corresponded to the river channel pattern transformation and was manifested by decreasing flow velocity (energy losses associated with bedform roughness and with the transportation of coarser particles). The flow velocity was generally greater in rivers of lithotype B3 and energy of sedimentary environment was more stable than during the deposition of lithotype B2.

## Introduction

Changes in fluvial styles throughout glacial-interglacial cycles were caused by combinations of periodic climate cooling and warming, ice sheet margin dynamics, fluctuations of the sea level and neotectonic movements (Kozarski 1983; Vandenberghe et al. 1994; Kasse et al. 1995; Huijzer and Vandenberghe 1998; Holbrook and Schumm 1999; Huisink 2000; Vandenberghe 2002, 2003; Van Huissteden and Kasse 2001; Houben 2003; Busschers et al. 2007; Erkens et al. 2009). The influence of the climate on fluvial processes has frequently been investigated by establishing variations in time of the pattern of river channels (a.o. Vandenberghe 1993, 1995, 2002, 2003; Mol 1997; Holbrook and Schumm 1999; Van Huissteden and Kasse 2001; Busschers et al. 2005, 2007, 2008; Starkel et al. 2007; Janssens et al. 2012; Meinsen et al. 2014); and by determining the changes in the textural and structural characteristics of fluvial sediments (a.o. Krzyszkowski 1996; Holbrook and Schumm 1999; Huisink 2000; Zieliński and Goździk 2001; Zieliński 2007). Changes of the riverbed morphology over the last glacial-interglacial transition have been found to be a direct result of spatial and temporal response to allogenic forcings (Schumm and Brakenridge 1987; Erkens et al. 2009). These forcings include climatic variations which influenced fluvial development by river regime changes, modification of the quality and extent of the vegetation cover in catchment and valley, the amount of the surface runoff and valley

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slope instability, limited infiltration due to permafrost, and of the type and intensity of the sediment supply to the river channel (e.g., aeolian sediments input and mass movements; a.o. Weckwerth and Pisarska-Jamroży 2015; Zieliński et al. 2015). Such climate-induced and vegetation-related transformation of river pattern (from a braided to a meandering system) were noted between the Late Pleniglacial and Late glacial periods in the alluvial reaches of mid-latitude European rivers (e.g., Vandenberghe et al. 1994; Kasse et al. 1995, 2005; Mol 1997; Busschers et al. 2007; Erkens et al. 2009; Janssens et al. 2012).

The principal indicators of the changes in the fluvial style are the changes in the morphology and the hydraulic parameters of the river channel (e.g., channel depth, width, river gradient), and the current energy (stream power). In the case of the Toruń Basin during the Weichselian glaciation the river regime depended on the extent of the nearby southern sector of the Scandinavian ice sheet (Weckwerth 2010, 2013; Weckwerth and Chabowski 2013; Pisarska-Jamroży 2015; Pisarska-Jamroży et al. 2015; Woronko et al. 2015) as well as to semi-cyclic alternations of climatic cooling and warming (references in the previous paragraph).

Recognition of the types of river channel patterns and their changes requires the reconstruction of qualitative and quantitative characteristics of the fluvial environment. The existing models of the river channel pattern transformation are based mainly on the relationships that may occur between the flow competences (Leopold et al. 1964; Schumm 1977, 1985) and relationships between the stream power and the particle sizes of the transported sediments (Van den Berg 1995; Thorne 1997; Van den Berg and Bledsoe 2003; Anisimov et al. 2008). It is quite well understood that the hydraulic parameters of the rivers (flow depth, hydraulic gradient, channel morphology, water flow velocity and stream power) affect the development of bedforms, to which fluvial lithofacies correspond (Ashley 1990; Southard and Boguchwal 1990; Southard 1991). The associations of bedforms built fluvial lithotypes, which in turn are relevant to recognizing and discriminating river channel patterns developed under the equilibrium channel state (Leopold 1964; Langbein and Leopold 1966; Schumm 1977, 1985; Knighton 1984; Van den Berg 1995; Kleinhans and Van den Berg 2011), and relating them to former discharge regimes.

Several previous geomorphological and geological studies of Weichselian outwash plain fluvial successions have had the objective to quantitatively characterize these environments, as does this study (a.o. Zieliński and Van Loon 1999, 2003; Blažauskas et al. 2007; Pisarska-Jamroży 2015). These past studies, however, did not consider to analyse the river regimes and their hydraulic parameters, in addition to changes in the ice-sheet edge position, also changed coevally due to climate fluctuations in the case of fluvial successions which nowadays are buried under Weichselian tills in

Poland (Krzyszkowski 1996; Wysota et al. 1996; Zieliński and Goździk 2001; Wysota 2002; Molewski 2007; Zieliński 2007; Weckwerth 2011). Thus, this paper bridges between the two types of insights here, sedimentological interpretation of buried fluvial sequences, and between the Late Weichselian superficial insights about the climate-derived transformation of river channel pattern.

As the fluvial succession filling up the Pleistocene Toruń Basin reaches a thickness of up to about 100 m, sufficient data became available for the reconstruction of the changes of the fluvial system that occurred between the successive phases of advance of the Scandinavian ice sheet. Such a reconstruction is the main objective of the present study, which focuses on the evolution of the climate-induced fluvial processes. Consequently, this paper concentrates on the sedimentology and paleohydraulic reconstructions, whereas earlier studies have been carried out describing the chronological information the fluvial formations described at the same sites (Weckwerth et al. 2011, 2013; Weckwerth 2013). The study had the following approach: recognition of the fluvial lithotypes and interpretation of the depositional mechanisms involved, assessment of the quantitative and qualitative features of the fluvial environment and construction of a model of the changes in the river channel pattern during the Weichselian glaciation that shows its controlling paleohydrological parameters and the sedimentary consequences.

## Geological setting

The Toruń Basin, located in the north-central part of Poland, belongs to the eastern part of the Toruń-Eberswalde ice-marginal valley (Fig. 1). During the Weichselian glaciation, this basin was fed by precipitation-dependent (periglacial) rivers flowing from the south, but just as well by streams flowing from the north that were fed by the meltwater from the ice sheet (Weckwerth 2013; Weckwerth and Chabowski 2013; Pisarska-Jamroży 2015; Pisarska-Jamroży et al. 2015). The position and extent of the basin was determined by valleys and glacial basins which evolved before the Last Glacial Maximum (LGM), i.e. before 24–19 ka ago (Makowska 1980; Wysota et al. 2009; Weckwerth 2010, 2013; Marks 2012).

The oldest valleys known in the area became infilled by fluvial sands and gravels which accumulated during the Holstenian interglacial (western part of Toruń Basin; Brykczyński 1986; Uniejewska and Nosek 1992; Marks 2005). Palynology records and results of OSL dating (Noryśkiewicz 1978; Makowska 1980; Wysota 2002) indicate that a next generation of river valleys evolved during the Eemian interglacial (Figs. 2, 3) when rivers bypassed parts of the Mesozoic basement in the area that had become uplifted by glacio-isostasy (Weckwerth 2013).

**Fig. 1** Location of the study area, showing the extent of the Scandinavian ice sheet during the Weichselian glaciation
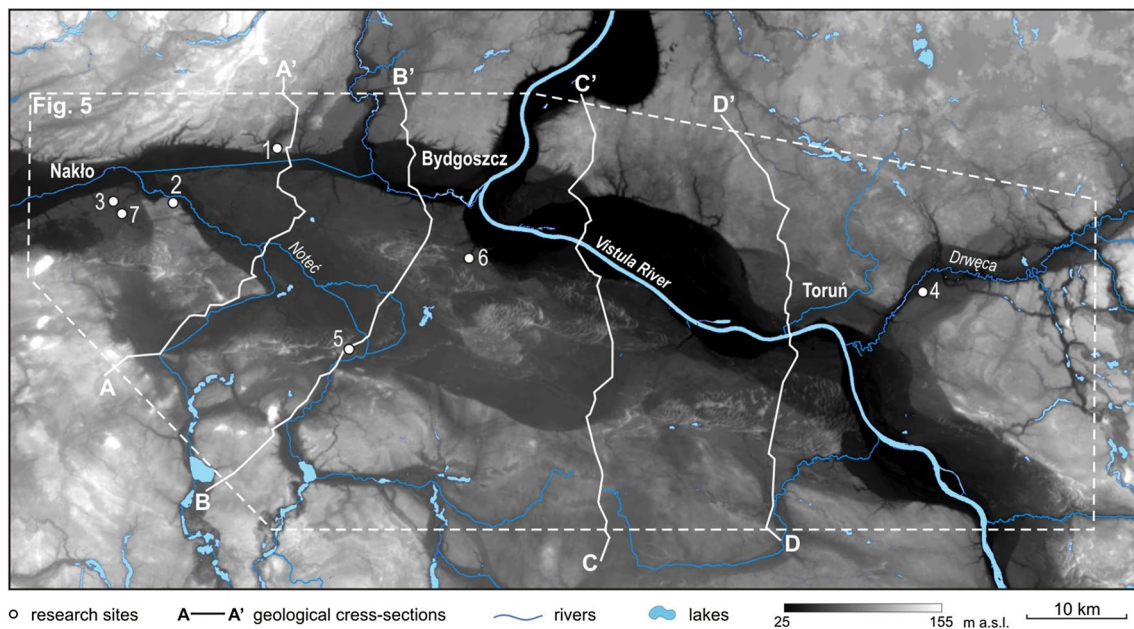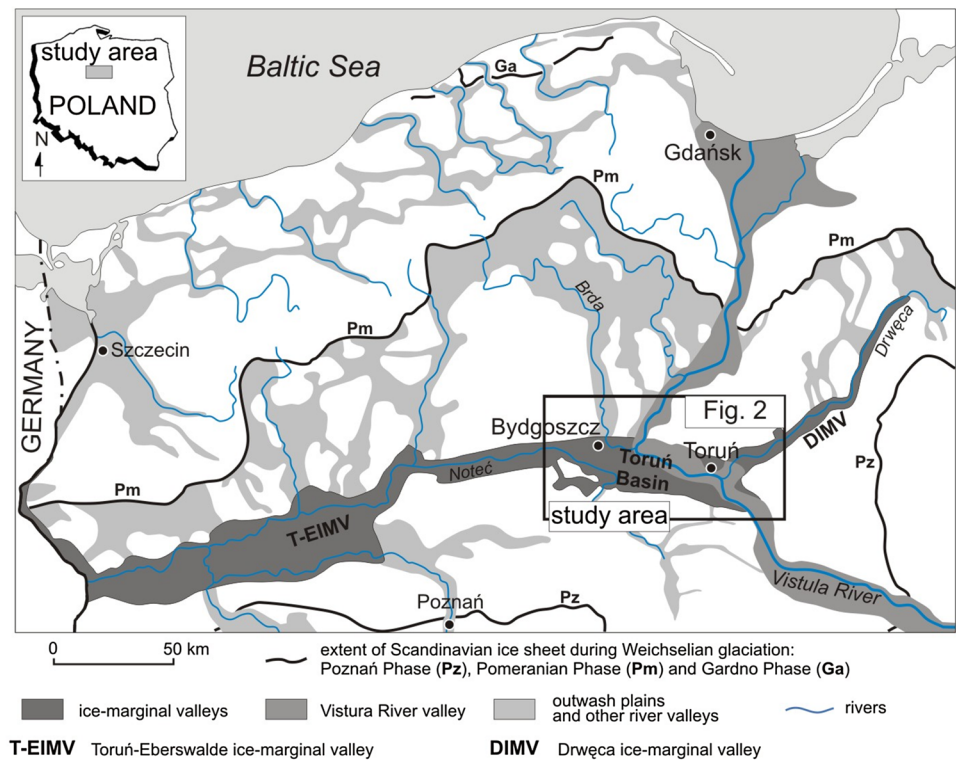


**Fig. 2** Relief of the Toruń Basin and the location of research sites and geological cross-sections (research sites: 1—Zielonczyn, 2—Chobielin, 3—Rozwarzyn, 4—Młyniec, 5—Nowe Dąbie, 6—Wypaleniska, 7—Paterek)

This glacio-isostatic rebound induced river incision at the end of Saalian glaciation (Termination II; postglacial transgression and sea-level stabilization; Marks 2005; Siddall et al. 2006) and was also a major cause of the development of Last-Interglacial valleys in areas of Mesozoic basement subsidence.

Weichselian fluvial and glaciofluvial successions are the fills of these Termination-II created valleys. They
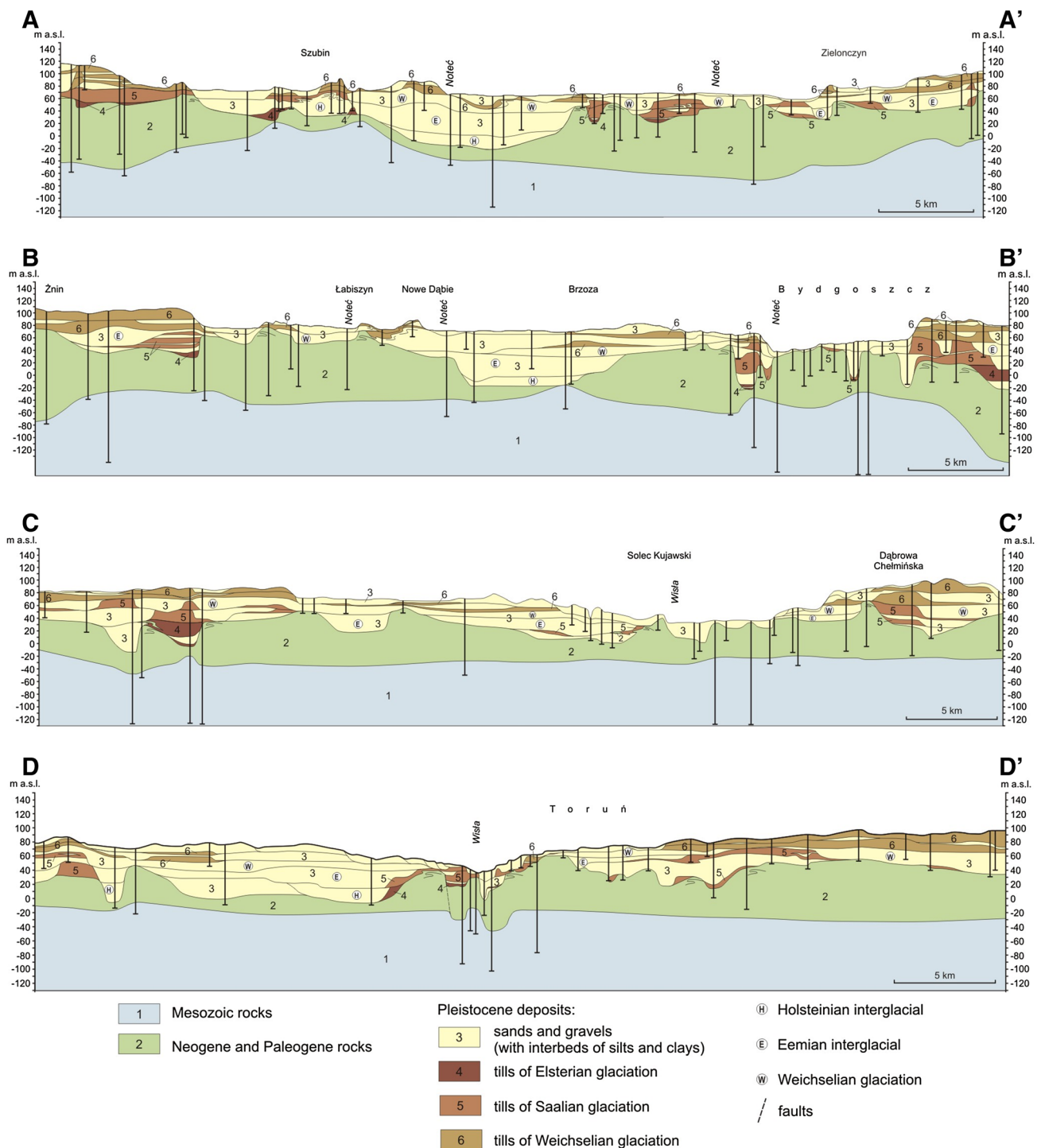
**Fig. 3** Geological cross-sections in the western part of the Toruń Basin (after Weckwerth 2010, 2013, modified)

constitute three fluvial formations separated by tills or by glaciolacustrine sediments (Fig. 4). The first fluvial formation is the Rzęczkowo Formation, which is over 29 ka years old (Wysota 2002; Wysota et al. 2009; Weckwerth et al. 2011). The two younger fluvial formations are the Zielonczyn Formation (deposited between 28

and 21 ka years ago) and the Noteć Formation, which includes deposits of the Toruń-Eberswalde ice-marginal valley (pradolina) that dates from 17 to 16 ka years ago on the base of geomorphological records and 14C dating (Galon 1961; Tomczak 1987; Wiśniewski 1990; Weckwerth et al. 2011; Weckwerth 2013). The lower members
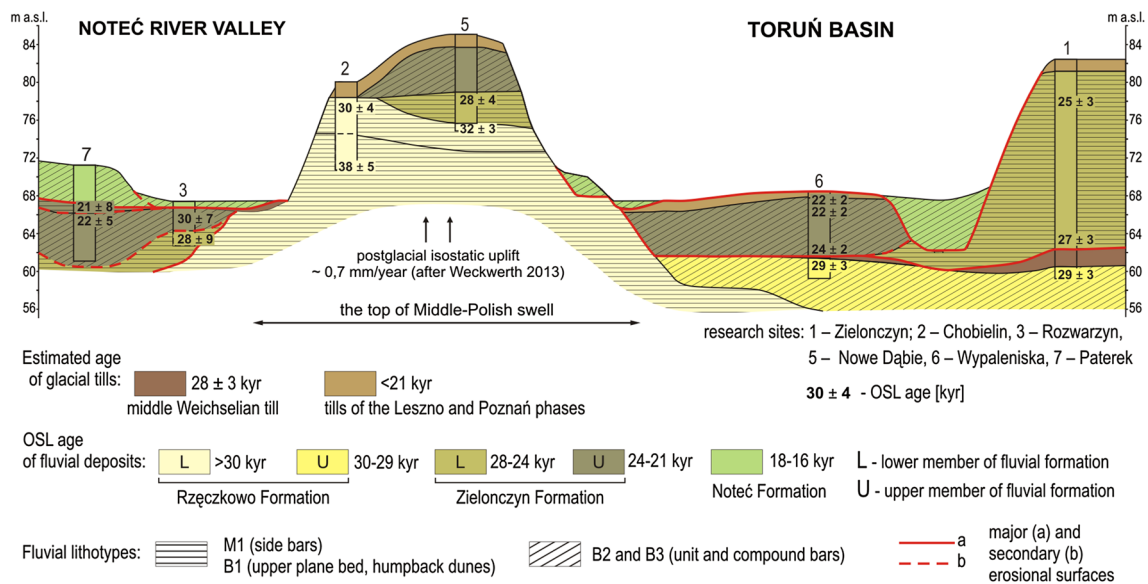
**Fig. 4** Synthetic lithological and stratigraphical cross-section of the Toruń Basin and Noteć River valley (after Weckwerth 2013, modified)
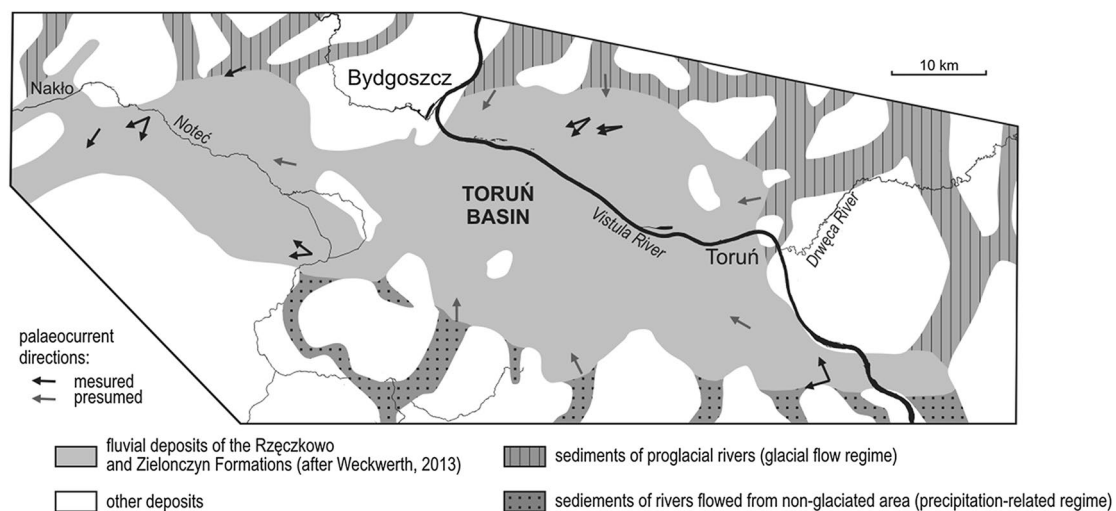


**Fig. 5** Toruń Basin and its fluvial system during the deposition of the Rzęczkowo and Zielonczyn Formations (after Weckwerth and Chabowski 2013, modified)

of the Rzęczkowo Fm. and Zielonczyn Fm. were accumulated in precipitation-fed rivers (Figs. 4, 5), whereas the sediments of their upper members were deposited by meltwaters (Wysota et al. 1996; Wysota 2002; Molewski 2007; Weckwerth 2013; Weckwerth et al. 2013). All members of the Rzęczkowo and Zielonczyn Formations were deposited in the Toruń Basin which was widened and deepened by glacial erosion and deformation during the Late Pleistocene (Weckwerth 2013). This large depression represents a sedimentary basin which merges

narrow valleys running from the south and tracks of meltwater streams flowing from the north (Fig. 5). This basin was drained to the west of Bydgoszcz through the antecedent river gaps which developed near Nakło due to glacio-isostatic rebound that induced river incision at the end of Saalian and Weichselian glaciations (Figs. 2, 5) (Weckwerth 2013).

**Table 1** Symbols of lithofacies and codes for channel bedforms and macroforms (after Miall 1978, 2006; Eyles et al. 1983; Zieliński 1995, 1998; Zieliński and Pisarska-Jamroży 2012; modified by the present author)

| Lithofacies code | | Lithogenetic code |
|---|---|---|
| Textural symbols | Structural symbols | |
| D—diamicton | m—massive | GL—glacial deposition |
| G—gravels | h—horizontal stratification, lamination | GB—gravelly bedforms and bars |
| S—sands | l—low-angle cross-stratification (small and medium scale) | SP—scour-pool infill, scouring channel |
| F—silt and/or clay | i—low-angle inclined stratification (large scale) | CC—chute channel |
| | t—trough cross-stratification | UP—upper plane bed |
| | p—planar cross-stratification (small and medium scale) | SB—sandy bedforms |
| | f—large scale planar cross-stratification | FM-u—transverse bars |
| | r—ripple cross-lamination | FM-c—transverse bars |
| | w—wavy lamination | LM—side bars, point bars |
| | e—scour infill | OF—overbank forms: crevasse splays and sand sheets |

## Methods

### Lithofacies analysis of fluvial sediments

Field data from seven key sites show that the fluvial deposits occur below the Weichselian tills and till-remnants. The conditions at which fluvial deposition took place were reconstructed on the basis of the structural and textural features. This made it possible to restrict the genetic origins of the various lithofacies to bedforms of specific channel types. In the studied exposures, several lithofacies coexist and form lithofacies associations related to specific fluvial sub-environments (Table 1). In turn, these lithofacies associations form groups of macroscale strata sets (cf. Bridge 1993), that we have related to the depositional sedimentary environments.

### Grain-size distribution

The grain-size distribution of the sands and gravels was determined by sieving at 1-φ intervals, whereas the silt and mud fractions were measured with a laser particle-size analyser (Analysette 22) at 0.25-φ intervals. The sediments were classified texturally according to the scale proposed by Udden and modified by Wentworth (1922). The statistical parameters of the grain-size distribution (median grain diameter, mean grain-size and sorting, skewness and kurtosis of the grain-size distribution) were assessed on the basis of the formulas by Folk and Ward (1957).

### Age control

Fluvial successions that compose lithostratigraphic unit, identified by lithological characteristics and stratigraphic position, distinct from other beds, were classified as formation which can be divided into members characterized by specific lithology (Tucker 2003; Marks et al. 2014). The ages of fluvial formation were obtained by OSL dating method, applying the single aliquots regenerative (SAR) dose protocol (Murray and Wintle 2000; Wintle and Murray 2006). The ages dataset was described and discussed in author's earlier studies under the terms of ages overestimation and reliability in paleogeographical context (Weckwerth et al. 2011, 2013; Weckwerth 2013).

### Paleohydraulic analysis

A series of paleo-flow parameters were quantified in the following manner. The river gradient ($S$) for the fluvial successions of the Rzęczkowo and Zielonczyn Formations was calculated using the Du Boy equation $\tau = \rho g D S$ and the Schields equation $\theta = \tau / \rho g R d_{50}$ (Paola et al. 1992; Julien and Wargadalam 1995; Dade and Friend 1998; Dade 2000; Paola 2000; Frings 2008; Weckwerth 2011), where $\tau$ is the bed shear stress ($kg m^{-1} s^{-2}$), $\rho$ is the water density ($kg m^{-3}$), $D$ is the hydraulic radius (mean flow depth for braided rivers) (m), $\theta$ is the dimensionless Shields parameter, $g$ is the gravitational acceleration ($= 9.81 ms^{-2}$), $R$ is the relative excess density of the sediment particles, and $d_{50}$ is the median grain diameter (m) (Fig. 6).

The flow depth $D$ (m) is related to the height of the bedforms and macroforms in a river channel (Simons and Richardson 1962), and thus proportional to the thickness of the individual lithofacies of planar and trough cross-stratified sands (Simons and Richardson 1962; Allen 1982; Weckwerth 2011; Zieliński 2014) (Fig. 6). The fluvial erosion effects were taken into account during the estimation of flow depth ($D$) because in braided rivers the mean preserved set thickness for bedforms is about 0.7–0.8 (cf. Paola and Borgman 1991; Mohrig et al. 2000). For massive or horizontally
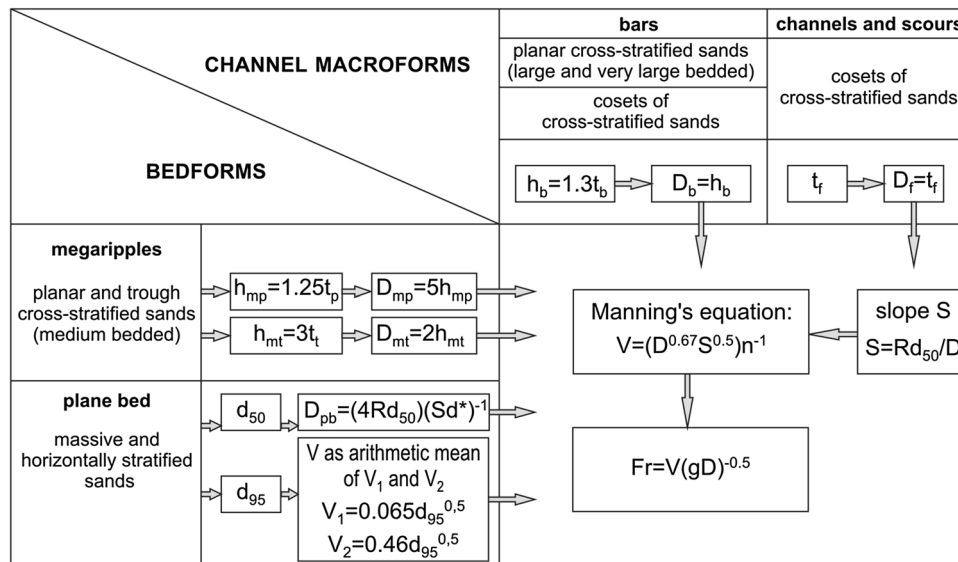
**Fig. 6** Scheme of paleohydraulic calculation principles for buried sedimentary facies of sand-bed-braided rivers. $t_p$—thickness of lithofacies of planar-stratified sands (medium bedded), $h_{mp}$—megaripple height calculated on the basis of the thickness of lithofacies of planar-stratified sands, $t_t$—thickness of lithofacies of trough-stratified sands (medium bedded), $h_{mt}$—megaripples height calculated on the basis of the thickness of lithofacies of trough-stratified sands, $D_{mp}$—mean flow depth calculated on the basis of the thickness of lithofacies of planar-stratified sands, $D_{mt}$—mean flow depth calculated on the basis of the thickness of lithofacies of trough-stratified sands, $t_b$—thickness of planar cross-stratified sands (large and very large bedded) or cosets of cross-stratified sands, $h_b$—transverse bar height, $D_b$—mean flow depth calculated on the basis of transverse bar height, $t_f$—thickness of channel and scour infill, $D_f$—mean flow depth calculated on the basis of thickness of channel and scour infill, $d_{95}$—95th percentile of grain-size distribution, $D_{pb}$—mean flow depth calculated for plane bed (after Julien and Raslan 1998), $d*$—dimensionless particle diameter (after Julien and Raslan 1998) [other symbols are explained in the text; $h_{mp}$ was calculated after Saunderson and Jopling (1980) and Zieliński (2014), $h_{mt}$ and $D_{mt}$ after Leclair and Bridge (2001), $D_{mp}$ after Zieliński (2014), $h_b$ after Paola and Borgman (1991), $D_b$ after Mohrig et al. (2000), $V_1$ and $V_2$ after Williams (1983), $S$ after Paola et al. (1992), Dade and Friend (1998), Dade (2000), Paola (2000) and Frings (2008)]

stratified sediments, the estimation of the flow depth was based on the median grain diameter ($d_{50}$) (Williams 1983; Julien and Raslan 1998). The results are consistent with the method that determines the flow depth on the basis of the lithofacies thickness (Leclair and Bridge 2001; Bridge 2003).

The average current velocity was estimated with Manning's equation, $V = (D^{2/3}S^{1/2})n^{-1}$, where n is Manning's coefficient (see Julien 2002). The Froude number was determined to compare with the sedimentological inferences and to establish an independent estimation of flow regime. This non-dimensional parameter was calculated using the equation $Fr = V(gD)^{-0.5}$ (Dey 2014) which shows the ratio of inertia forces to the gravity forces (Fig. 6). The sinuosity of the river channel was estimated using the equation $SN = 1 - (VR/252)^{-2}$, where SN is the channel sinuosity and VR is the maximum angular range of the mean azimuth of the crossbeds (Langbein and Leopold 1966; Miall 1976).

# Sedimentological characteristics of the fluvial deposits

## Sediments of the lower members of the Weichselian fluvial formations

### Fluvial succession at the Chobilelin site

Sediments which represent the lower member of the Rzęczkowo Fm. are present at the Chobielin site and were deposited between $38.3 \pm 4.7$ and $29.5 \pm 3.9$ ka ago (Figs. 4, 7). These are 9 m thick and consist of three similar lithofacies associations—Si(SFw), Si(Sm,Sr) and Si(Sm,Sr,Src,Se)—in which a large-scale and low-angle inclined stratification of fine-grained sands (Si) dominate (Fig. 7). The laminae of lithofacies Si dip up to 8° to the south and the south-east. Sands and silts with wavy lamination (SFw) and ripple cross-lamination (Sr) and beds with climbing ripples: (lithofacies Src), scour infillings (Se) and massive sands (Sm) are also present in this fluvial succession. All the mentioned lithofacies form fining-upward sequences Sm→Si→Sr(Src)/Se.
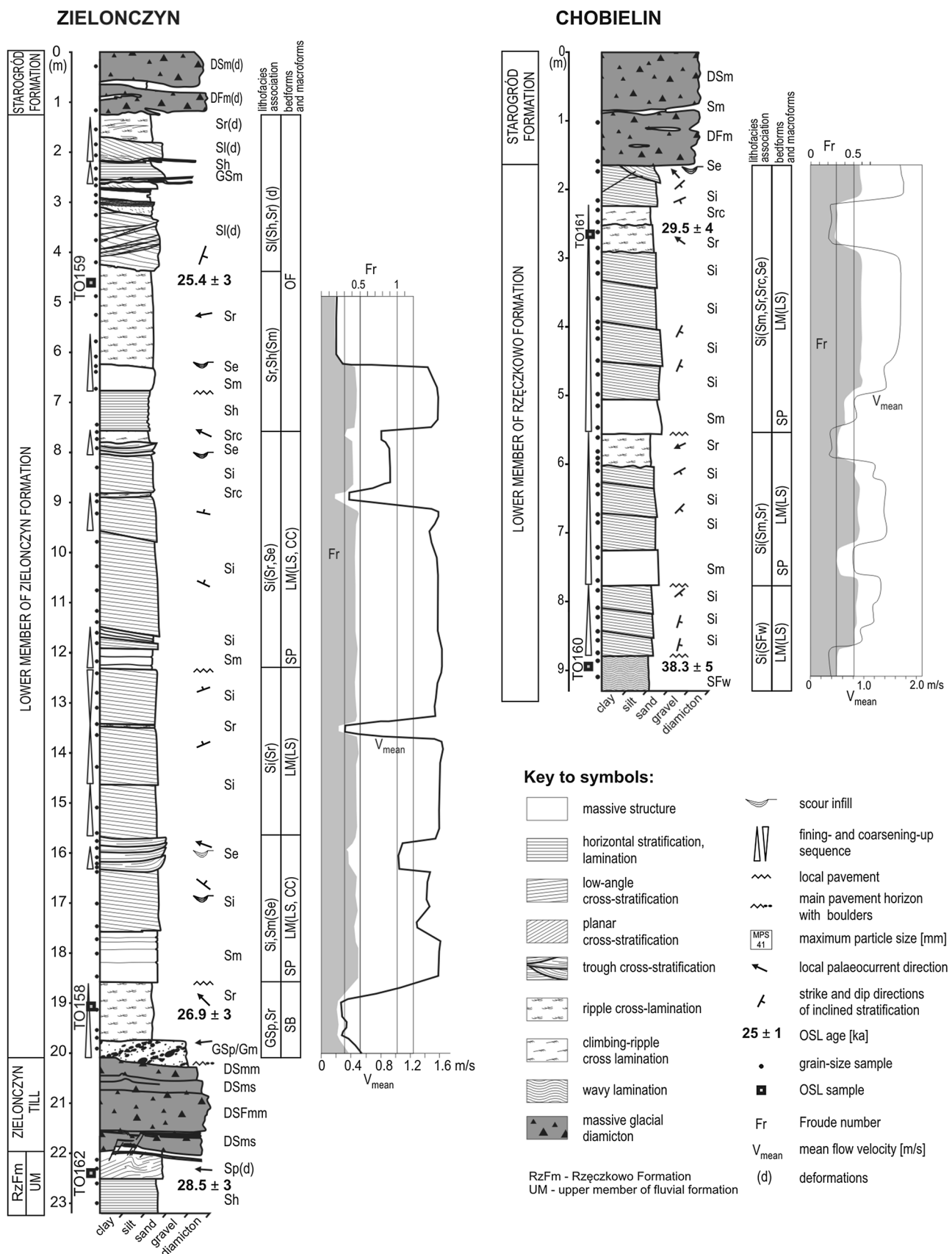
**Fig. 7** Lithofacies log of the Zielonczyn and Chobielin sites (explanation of lithofacial and lithogenetic codes in Table 1)

**Fig. 8** Lithofacies log of the Rozwarzyn, Młyniec and Nowe Dąbie sites (for legend, see Fig. 7 and Table 1)

The fluvial succession at Chobielin can be divided into two parts on the basis of the grain-size distribution. The lower part occurs at the depth of 9–5.6 m and consists of well-sorted fine-grained sands, in which the average value of median grain diameter ($d_{50}$) is 0.14 mm. The upper part, between 5.6 and 1.7 m, is built of moderately sorted fine-grained sands in which the median grain diameter ($d_{50}$) is about 0.19 mm.

### Fluvial succession at the Zielonczyn site

Sediments of the lower member of Zielonczyn Fm. are exposed at the Zielonczyn site and reach a thickness of 19.5 m. These were deposited between $26.9 \pm 3$ and $25.4 \pm 3$ ka ago (Figs. 4, 7) and consist of six lithofacies associations (Fig. 7). The basal part of the first lithofacial association (GSp,Sr) is built of cross-stratified and poorly sorted sandy gravels (lithofacies GSp) in which the average median grain diameter ($d_{50}$) is 1.77 mm. These sediments are overlain by ripple cross-laminated and well-sorted fine-grained sands (lithofacies Sr) in which the average median grain diameter ($d_{50}$) is 0.15 mm. The grain-size distributions of lithofacies association GSp,Sr are symmetric and meso- or platykurtic. The next lithofacies associations [Si,Sm(Se), Si(Sr) and Si(Sr,Se)] form three similar sequences Sm→Si→Sr/Se. Their lower segments are composed of massive fine-grained sands (Sm), in which median grain diameters ($d_{50}$) vary between 0.15 and 0.16 mm. Middle segments consist of sets of large scale and low-angle inclined stratification of fine-grained sands (Si) in which laminae dip at 6°–14° to the northeast, southeast and southwest (Fig. 7). Sediments of lithofacies Si are normally graded (the median grain diameters decrease from 0.17 to 0.15 mm) and well or moderately sorted. The uppermost segments of each sequences comprise rippled fine-grained sands (Sr) in which the median grain diameters differ between 0.13 and 0.16 mm. These segments may contain sandy scour infillings of a depth up to 0.4 m (lithofacies Se, median grain diameter $d_{50}$ between 0.16 and 0.18 mm).

The upper part of the fluvial succession at the Zielonczyn site is composed of two lithofacies associations Sr,Sh(Sm) and Sl(Sh,Sr)(d) (Fig. 7). The first is dominated by ripple cross-laminated fine-grained sands (Sr) the thickness of which is 2 m. These sediments lie on the horizontally stratified sands (Sh) or on massive sands (Sm). The second lithofacies association [Sl(Sh,Sr)(d)] comprises lithofacies of medium-scale and low-angle cross-stratified fine-grained sands (Sl), which are interbedded with ripple cross-laminated sands (Sr) and with horizontally stratified sands (Sh). Their sedimentary structures were deformed by the development of reverse faults and clastic dykes.

### Fluvial deposits at the Rozwarzyn, Młyniec and Nowe Dąbie sites

Fluvial deposits of the lower members of Weichselian fluvial formations were also found at the Rozwarzyn, Młyniec and Nowe Dąbie sites (Fig. 8). These sediments are 5 m thick and contain three lithofacies associations at the Rozwarzyn site. The first of them [lithofacies association Sh,Sl(Sr,Sp)] is dominated by middle- and small-scale beds of horizontally stratified and low-angle cross-stratified sands (median grain diameters $d_{50}$ are in the range of 0.15–0.16 mm). These sediments are well sorted and have symmetric or positively skewed grain-size distributions. Locally, they are interbedded with ripple cross-laminated sands and with small- and middle-scale lithofacies of planar cross-stratified sands (Sp) which are coarser (average $d_{50} = 0.22$ mm) and moderately sorted. Additionally, normally graded, large-scale and low-angle cross-stratified sands (Sl) occur in the upper part of the lithofacies association Sh,Sl(Sr,Sp). This lithofacies is characterized by sigmoidal crossbeds and varied thicknesses (from 0.7 up to 2.5 m).

The lower members of the Weichselian fluvial successions are represented by lithologically similar deposits at the Młyniec and Nowe Dąbie sites (Fig. 8). Sands showing medium-scale and small-scale planar cross-stratification (Sp) and rippled fine-grained sands were encountered in both sites; scour infillings additionally occur (lithofacies Se). These deposits are moderately well and moderately well sorted, while their median grain diameter ($d_{50}$) varies from 0.15 up to 0.29 mm. Moreover, massive and matrix-supported sandy gravels (lithofacies GSm) were recognized at the Młyniec site, whereas the lithofacies association Sr(Src,SFh) was observed at Nowe Dąbie site. Sandy beds of the ripple-drift cross-lamination occur commonly in this lithofacies association. Additionally, horizontally laminated silty sands (lithofacies SFh) were identified at a depth of 7.2–7.6 m (Fig. 8).

### Sediments of the upper members of the Weichselian fluvial formations

The upper member of the Zielonczyn Formation at the Nowe Dąbie site contains two lithofacies associations Sp(Sr,Se) and Sf,Sp(Sl,Sr,Sh) (Fig. 8). The first of them is 2.2 m thick and comprises medium-scale planar cross-stratified sands (lithofacies Sp, profile Nowe Dąbie 2; Fig. 8). These reverse graded sediments (median-grain diameter $d_{50}$ increases from 0.29 up to 0.34 mm) are moderately well and well sorted and interbedded with scours infillings (lithofacies Se) or with ripple cross-laminated fine-grained sands (Sr). Lithofacies association Sf,Sp(Sl,Sr,Sh) has a thickness of 4.75 m. Its lower part is dominated by fine- and medium-grained sands of inclined stratification (lithofacies Sf, median-grain
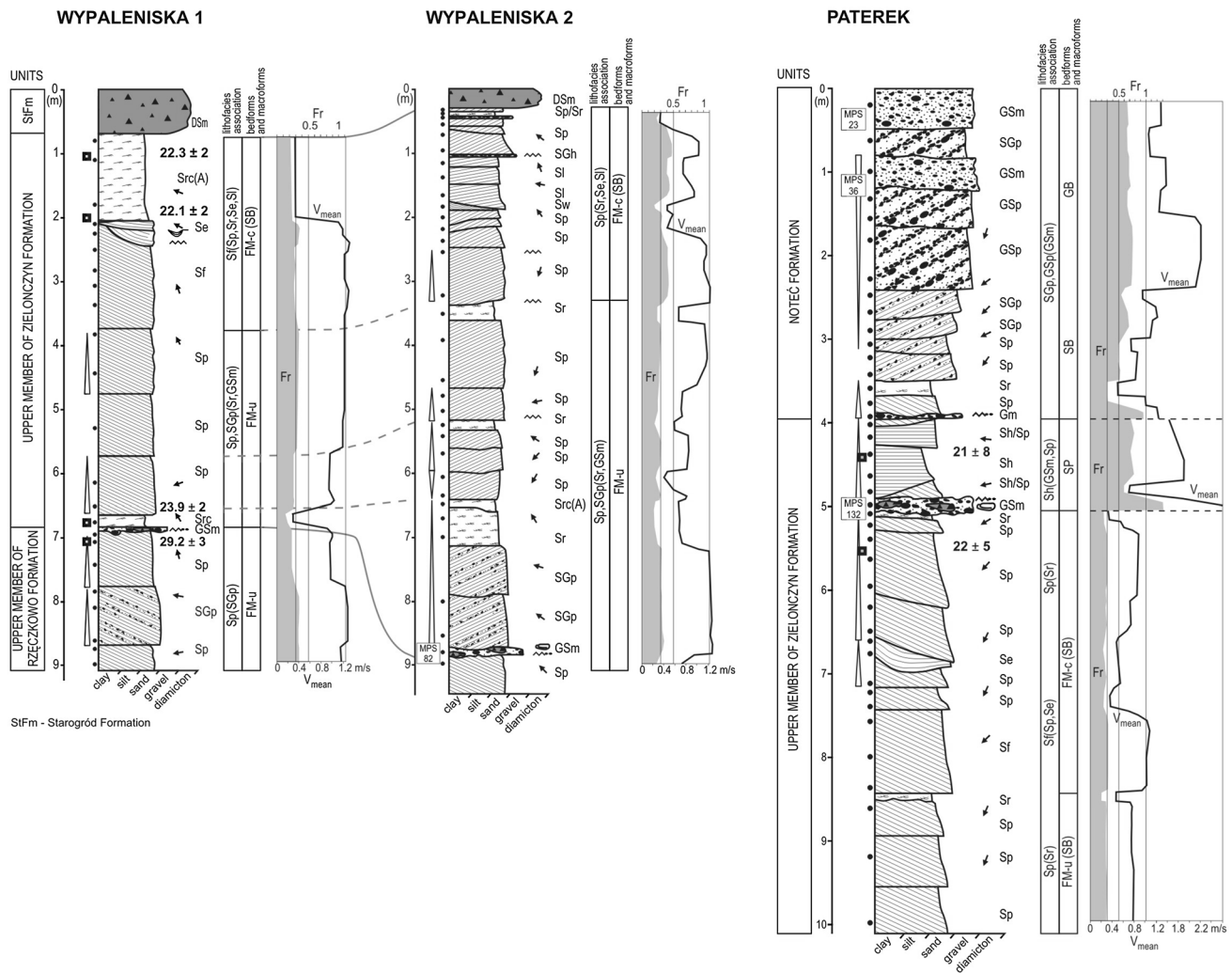
**Fig. 9** Lithofacies log of the Wypaleniska and Paterek sites (for legend, see Fig. 7 and Table 1)

diameter $d_{50}$ from 0.16 up to 0.31 mm) which reach a thickness up to 2.1 m and are well and moderately sorted (profile Nowe Dąbie 1; Fig. 8). These sediments are overlain by low-angle cross-stratified sands (Sl), sands of scours infills (Se) and by small-scale planar cross-stratified sands (Sp) with interbeds of horizontally stratified sands (Sh). Similar large-scale sandy beds of planar and inclined stratification (lithofacies Sp and Sf, respectively) were recognized in the upper part of the fluvial succession at the Rozwarzyn site (Fig. 8).

Fluvial sediments at the Wypaleniska site represent two Weichselian fluvial formations which were deposited up to $29 \pm 3$ ka ago (Rzęczkowo Fm.) and between 24 and 22 ka ago (Zielonczyn Fm.; Weckwerth et al. 2011; Weckwerth 2013). The first of them (Rzęczkowo Fm.) consists of medium- and fine-grained planar cross-stratified sands with an admixture of granules and pebbles [lithofacies association Sp(SGp), Fig. 9]. These deposits are normally graded (median-grain diameter $d_{50}$ decreases from 0.40 to 0.17 mm)

and moderately well and well sorted, while their grain-size distributions are negatively skewed and leptokurtic.

Sediments of the Zielonczyn Fm. are represented by two lithofacies associations Sp,SGp(Sr,GSm) and Sf(Sr,Se,Sl) at the Wypaleniska site (Fig. 9). The first lithofacies association comprises cosets of medium-scale lithofacies of planar cross-stratified fine- and medium-grained sands (median-grain diameter $d_{50}$ changes from 0.17 to 0.39 mm, profile Wypaleniska 1) or two-part sequences Sp→Sr which consist of planar cross-stratified sands (Sp) with cappings of rippled and fine-grained sands (Sr) (profile Wypaleniska 2). Sands of planar cross-cosets are moderately well and moderately well sorted, whereas their grain-size distributions vary from negatively to very positively skewed. The second lithofacies association [Sf(Sp,Sr,Se,Sl)] is dominated by very thickly planar cross-bedded fine-grained sands (median-grain diameters vary between 0.21 and 0.32 mm; Fig. 9). These sediments are overlain by medium- and small-scale planar

**Fig. 10** Typical Weichselian fluvial lithofacies associations and fluvial styles in relation to the flow regime conditions (changes of Froude number) and sediments size ($d_{50}$ variations) reflecting river channel pattern transformations (explanation of lithofacial and lithogenetic codes in Table 1)
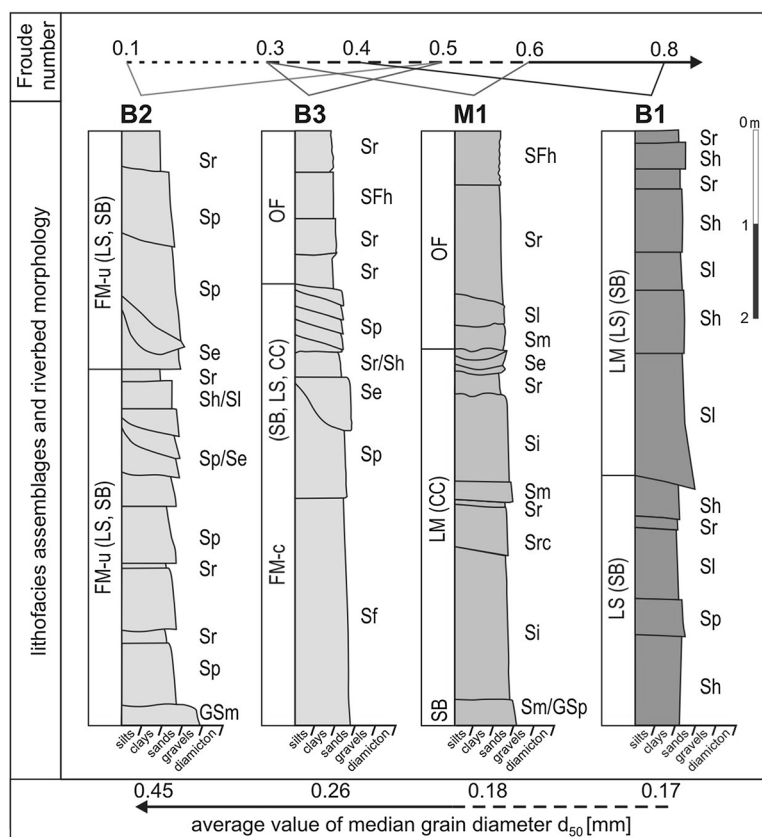


cross-stratified sands (Sp), sandy scour infillings (Se), rippled sands (Sr) and low-angle cross-stratified sands (Sl).

A similar fluvial succession was recognized at the Paterek site, where the sandy cosets of mid- and large scale planar cross-stratification (lithofacies Sp and Sf, respectively),



**Fig. 11** Channel depth in relation to water flow velocity and flow regime (Froude number) for Weichselian fluvial lithotypes

rippled sands (Sr) and sandy scour infillings (Se) form two-part sequences Sp→Sr and Sf→Se/Sp. Their lower segments (lithofacies Sp and Sf) comprise medium-grained and moderately well-sorted sands (median-grain diameters $d_{50}$ are in the range 0.40–0.54 mm) which are characterized by positively and very positively skewed grain-size distribution. The upper segments consist of small- and medium-scale lithofacies Sp, Se and Sr in which fine-grained sands predominate (median-grain diameters $d_{50}$ vary from 0.15 up to 0.40 mm).

The upper member of Zielonczyn Fm. at the Paterek site is built of lithofacies association Sh(GSm,Sp). It's basal part contains massive pebbles with a sandy matrix (lithofacies GSm, average median-grain-diameter $d_{50}=8.15$ mm) (Fig. 9). These sediments are overlain by horizontally stratified fine-grained sands (Sh) which are normally graded (median-grain-diameter $d_{50}$ decreases from 0.24 to 0.17 mm) and moderately sorted. These are interbedded with small-scale planar cross-stratified and medium-grained sands (lithofacies Sp, average median grain diameter $d_{50}=0.41$ mm).
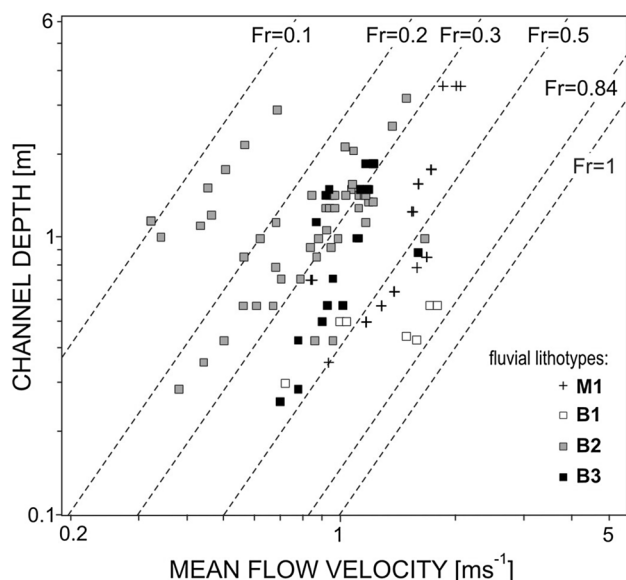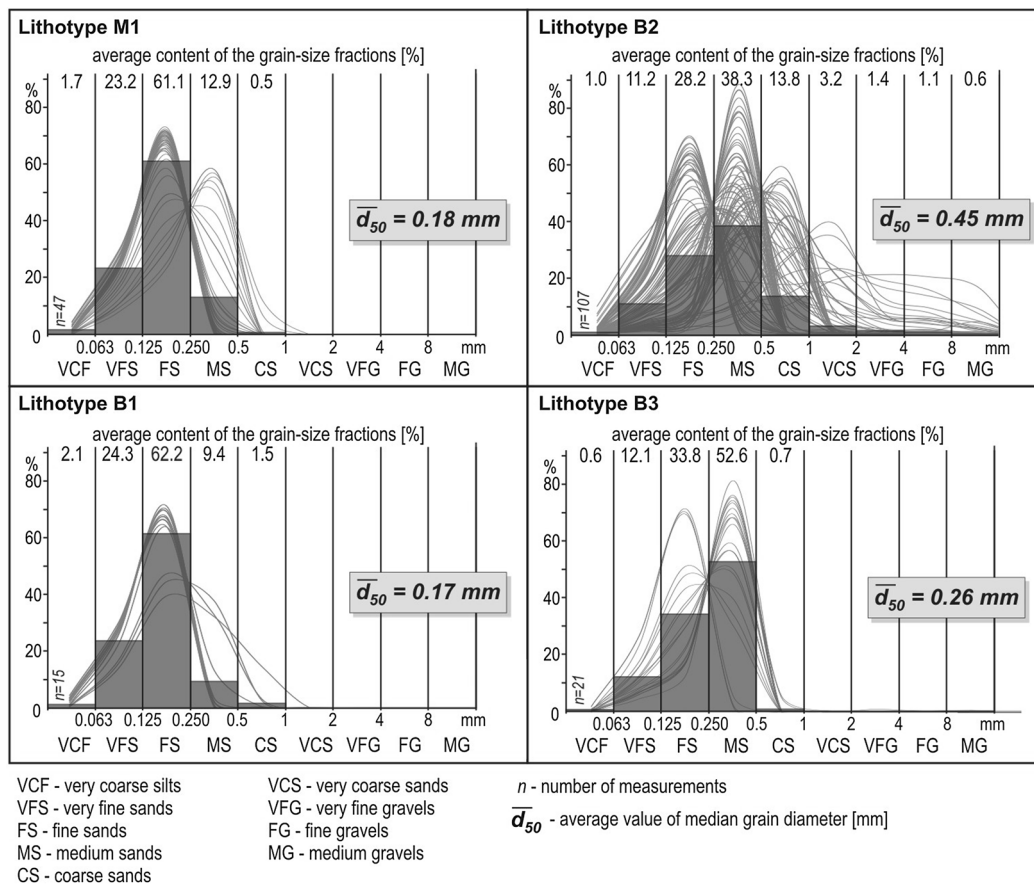
**Fig. 12** Grain-size compositions of Weichselian fluvial lithotypes in the Toruń Basin

## Fluvial lithotypes, riverbed morphology and channel patterns

The sedimentological results of Weichselian fluvial successions in the Toruń Basin allowed for four fluvial lithotypes to be recognized: one of a meandering-style sand-bed river type (M1), and three of braided-style sand-bed river styles (B1, B2 and B3). These lithotypes relate to the sand-bed rivers of different fluvial style.

## Lithotype of a sand-bed, high-energy meandering river (M1)

The lithotype M1 is represented by lithofacies association Si(Sm,Sr,Se) (Chobielin and Zielonczyn sites; Figs. 7, 10) which are arranged into stacked sequences Sm→Si→Sr(Se). The indicative large-scale lithofacies of low-angle inclined stratification of fine-grained sands (Si) capped by rippled sands (Sr) and sandy scours infillings (Se), represent point-bar sequences (Tunbridge 1981, 1984; Stear 1983; Miall 1985, 2006; Olsen 1988, 1989; Julien 2002; Bridge 2003). The surfaces of point-bars sloped at up to 8° in the case of a

shallow channels (depth up to 1 m) and at 8°–14° if channels reached a depth of 1–3.5 m. These low inclined point-bar surfaces developed in channels of 20–30 m wide, the sinuosity of which varied from 1.25 up to 1.5 (see Leeder 1973).

The deposition of fluvial lithotype M1 took place mostly under transitional water flow conditions from subcritical to supercritical or under supercritical flow. These conditions were related especially to the high water flow velocity (up to 1.7 ms$^{-1}$), whereas the Froude number reached a maximum only of 0.6 (Fig. 7,11). Similar low values of the Froude number (0.16–0.68) for supercritical flow were observed in modern sand-bed braided and meandering rivers (Leopold et al. 1964; Karim 1995; Julien and Raslan 1998; Prent and Hickin 2001). Moreover, the experimental results (Cheel 1990b; Julien and Raslan 1998; Best 2005; Fourrière et al. 2010) reveal that the supercritical flow may occur in a wide Froude number range 0.21–1.6. These true low values of the Froude number, experimentally obtained for a sequence of which the sedimentology indicates conditions of supercritical flow, may reflect the high rate of fine-grained sands suspended in a water column (Bridge and Best 1988; Baas and Best 2002; Best 2005). Moreover,

the average median grain diameter ($d_{50}$) of fluvial lithotype M1 is 0.18 mm (Fig. 12). Low-angle cross-stratification and parallel lamination (inclined or horizontal) commonly occur in these fine-grained sands, as was reported usually in sands characterized by median grain diameter in the range 0.1–0.3 mm and transported under the condition of the Froude number lower than 0.84 (Ashley 1990; Southard and Boguchwal 1990; Southard 1991).

The high transport rate of fine-grained sands of the fluvial lithotype M1 caused suppression of turbulence in the moving bedload dispersion (see Allen and Leeder 1980). Thus, the stable upper-plane bed and inclined parallel stratification of point-bar sequences (lithofacies Si) were formed due to the migration of low-relief bed waves of up to 20–25 mm high and 1.3 m long (see Bridge and Best 1988; Paola et al. 1989; Best and Bridge 1992). Similar supercritical water flow conditions were responsible for the deposition of massive sands (lithofacies Sm; Chobielin site; Figs. 7, 10) (see Alexander et al. 2001; Duller et al. 2008) as a result of scour-pool infilling (Gorrell and Shaw 1991; Russell et al. 2007) in a river thalweg. Moreover, two-dimensional dunes migrated and sandy-gravelly upper flow regime sheets developed in this riverbed section (lithofacies GSp and SGm, respectively; Zielonczyn site; Fig. 7) (Fielding 2006). The rippled sands (Sr) and sandy chute cut-off infillings (Se) which form the upper segments of point-bar sequences, are indicators of waning flow conditions at the end of flood events and reflect episodes of sedimentation and erosion (Farrell 1987; Zieliński 1992, 1993; Miall 2006).

The depositional environment of the fluvial lithotype M1 and its high-energy nature show similarities to the sedimentary environment of flashy, ephemeral, sheet flood, sand-bed braided river (Biju Creek type; see McKee et al. 1967; Tunbridge 1981, 1984; Miall 1985, 2006). However, the common occurrence of the low-angle inclined stratification of fine-grained sands (large scale lithofacies Si) indicates lateral accretion of point-bars developed in an ephemeral stream (Stear 1983). Thus, the fluvial lithotype M1 recognized in the Toruń Basin corresponds to the sedimentary environment of an ephemeral sandy meandering river (Miall 2006).

### Lithotype of a shallow, high-energy sand-bed braided river (B1)

Sediments of the fluvial lithotype B1 are similar to the lithotype M1 in terms the textual features (Fig. 12). These deposits consist mainly of fine-grained sands (61–62%), while the average value of the median grain diameter is 0.17–0.18 mm. The key indicator of the lithotype B1 is the dominance of horizontally stratified fine-grained sands

(lithofacies Sh) (Rozwarzyn site and the upper member of the Zielonczyn Fm. at the Paterek site; Figs. 9, 10). These lithofacies represent sheet-like beds veneering the floor of a sand-bed-braided river under the condition of supercritical flow (flow velocity up to 1.4 ms$^{-1}$; Figs. 9, 11). The deposition within shallow streams was related to a wide sheet flow rather than channelised flow (flow depth 0.4–0.7 m) and took place at the low Froude number as a consequence of the high concentration of suspended sediments (Bridge and Best 1988; Cheel 1990a; Julien and Raslan 1998; Best 2005; Fourrière et al. 2010). Thus, the association of sediments in suspension with supercritical flow conditions implies abundant horizontally stratified sands (see Bridge and Best 1988; Paola et al. 1989; Southard 1991; Best and Bridge 1992; Alexander et al. 2001; Fourrière et al. 2010). These lithofacies were formed by the migration of antidunes and low-relief bed waves.

The upper segment of the lithotype B1 (large-scale and low angle crossbeds Sl with sigmoidal laminae) reflects lateral aggradation of humpback dunes (Kostaschuk and Villard 1996; Mohrig and Smith 1996; Kostaschuk 2000; Winsemann et al. 2011). The morphology of these bedforms was changeable due to a variable rate of suspended sediment concentration (Kostaschuk and Villard 1996; Mohrig and Smith 1996; Kostaschuk 2000; Lang and Winsemann 2013). Therefore, the deposition of fine-grained sands from the suspended load close to the bed controls the lee side of the slope angle (Kostaschuk et al. 2009). Additionally, the migration of humpback dunes that were formed around the transition to a supercritical flow could be considered as the initial phase of the riverbed transformation from upper-plane beds to the dunes of the lower flow regime (Fielding 2006).

Sedimentological features of the fluvial lithotype B1 are related to the Bijou Creek fluvial lithotype (McKee et al. 1967; Williams 1971; Tunbridge 1984) and to the model of the flashy, ephemeral, sheet flood, sand-bed river (distal, sheet flood braided model of Miall 1985). A similar fluvial succession was described by Wysota (2002) based on his studies of the Rzęczkowo Formation in the area of the Lower Vistula Valley (in the northern vicinity of the Toruń Basin).

### Lithotype of a sand-bed braided river with unit bars (B2)

The third fluvial lithotype B2 contains lithofacies association Sp(Sr,Se,Sl,Sh) which is dominated by planar or trough cross-stratified sands (lithofacies Sp and St, respectively; Figs. 9, 10). Typical lithofacies (Sp) are interpreted as related to the unit transverse bars and numerous sandy megaripples which covered the bed of a braided river (Smith 1978; Miall 1985; Bridge 1993, 2003). Sedimentation took place under the transitional conditions between the lower

and upper parts of the lower flow regime (Froude number Fr 0.3–0.4; Figs. 9, 10, 11). The surfaces of the unit bars were dissected by small cross-bar channels with sandy infillings (Se) and covered by sands of diminished dunes (Sl). During the waning flood stages, fine-grained sands were accumulated due to the tractional deposition under the condition of a ripple-phase bed (lithofacies Sr).

The fluvial lithotype B2 was deposited from a low-energy flow characterized by a wide range of flow depth (0.3–2.8 m) and velocity of up to $1.6 \text{ ms}^{-1}$ (Figs. 9, 11). Its maximum value occurred during the deposition of massive gravelly sandy sheet-like beds (GSm). The dominance of crossbeds (lithofacies Sp) and minor sequences of laminated sand sheets, rippled sands and infillings of cross-bar channels are the common features of a "Plate-type" sand-bed braided river (shallow perennial braided; Miall 1985, 2006).

## Lithotype of a sand-bed-braided river with compound bars (B3)

The lithotype B3 contains two lithofacies associations representing two fluvial sub-environments (Fig. 10). The main association Sf(Sp,Sl,Se,Sr) was deposited within the channel of sand-bed-braided river and is dominated by large-scale planar cross-stratified sands (Sf) formed by the downstream migration of macroforms lee side (Nowe Dąbie, Wypaleniska and Paterek sites; Figs. 8, 9, 10) (Williams 1971; Cant and Walker 1978; Miall 1978, 1985; Miall and Jones 2003). This type of lithofacies predominantly occurred at the depths of 1.5–2.5 m below the surface of compound bars, which corresponds to those recognized in the South Saskatchewan River (Cant and Walker 1978; Sambrook Smith et al. 2006; Ashworth et al. 2011). These macroforms develop in a sand-bed-braided river due to the unit transverse bar accretion and were also observed in the Platte (Blodgett and Stanley 1980; Allen 1983; Crowley 1983; Bridge 1993, 2003), Vistula (Babiński 1992), Yamuna (Best et al. 2003) and Niobrara (Skelly et al. 2003) rivers. Moreover, lithofacies Sf may well be interpreted as an effect of alternate bar migration (McCabe 1977; Crowley 1983), the crest of which was dissected by chute channels (lithofacies Se). Alternate bars may be subsequently subjected to transformation into midchannel compound bars (see Bridge 1985, 2003; Bridge et al. 1986; Babiński 1992).

The upper part of the compound bar succession was accumulated during the flood as a result of two-dimensional dunes and ripple migration (lithofacies Sp, Sr, respectively) which constituted sand shoals on the compound bar surface (Sambrook Smith et al. 2006; Ashworth et al. 2000, 2011). These bedforms may change rapidly into the diminished dunes (lithofacies Sl) or into upper plane beds (lithofacies Sh) due to the predominance of fine-grained sands transported within the stream under the condition of increasing

water flow velocity caused by the decrease of the flow depth (see Ashley 1990; Southard 1991). Thus, variations of hydraulic parameters from the lower to upper flow regime tended to change the morphology of the compound bar surfaces in abundant areas of sand flats in sand-bed-braided river. Moreover, the compound bars were dissected by small cross-bar channels filled with sands (lithofacies Se) during the falling flood stage (Bridge and Lunt 2006).

The secondary lithofacies association Sr(Src,SFh) of the fluvial lithotype B3 represents the overbank sub-environment where floodplain fines were accumulated as a result of overbank splay development and sheet flows (Fig. 10) (Bridge 2003; Bridge and Lunt 2006; Miall 2006).

Sediments of the fluvial lithotype B3 were deposited predominately under transitional conditions between the lower and upper parts of the lower flow regime ($Fr \approx 0.3$), although, there were transitional conditions from subcritical to supercritical flow in the zones of upper stage plane beds and in cross-bar channels (Figs. 10, 11). The water flow velocity was variable (up to $1.6 \text{ ms}^{-1}$).
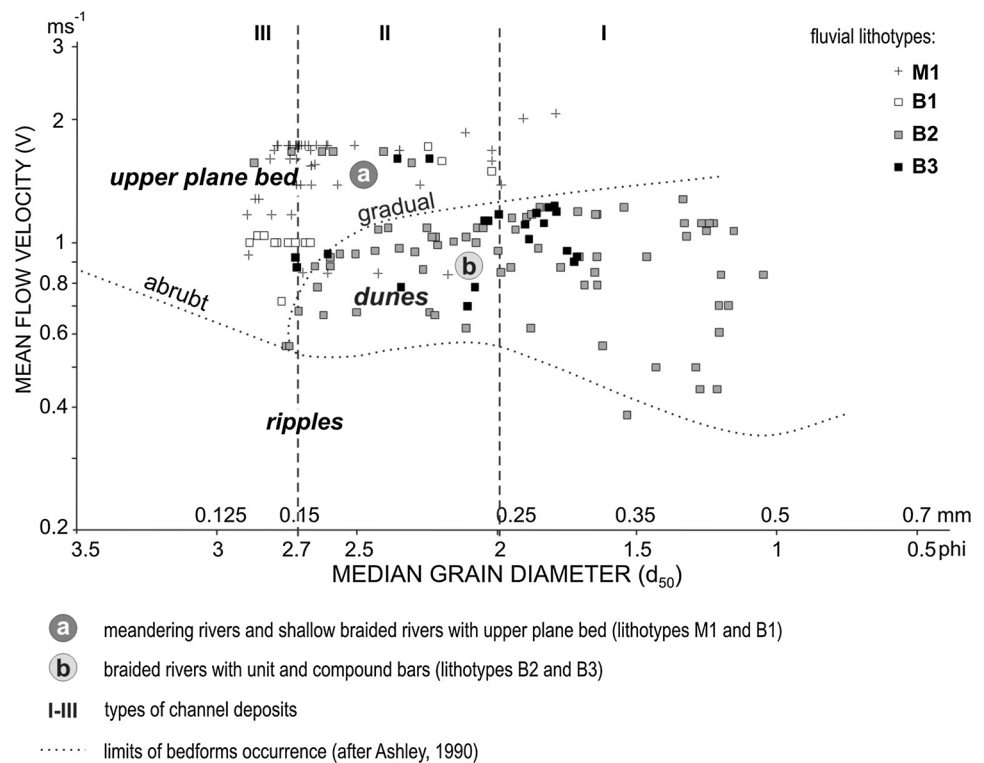
Qualitative features of the sedimentary environment of the fluvial lithotype B3 are similar to those in the South Saskatchewan River (Cant and Walker 1978) and are comparable with the depositional environment of deep perennial sand-bed braided river (Miall 1985, 1993, 1994, 2006) and with the lithotype D-3 described by Zieliński (1993) as sediments of aggrading-braided channels dominated by foreset bar accumulation.

## The relations between fluvial deposits and hydraulic conditions of the bedform development

Hydraulic parameters of the rivers (e.g., water flow velocity) affect the development of bedforms, to which fluvial lithofacies correspond. Due to this the relationships between properties of fluvial sediments (structural and textural) and flow competence reflect hydrodynamic conditions of the fluvial depositional environment. For this reason, quantitative interpretation of sedimentary structures and textural properties of fluvial lithotypes is fundamental.

Lithotypes M1 and B1 are characterized by the most similar grain-size distributions (Fig. 12) which are generally symmetrical or—locally—show a negative skewness associated with an admixture of fractions coarser than average. Fine-grained sands represent 61–62% on average, while the average values of the median grain size distributions are 0.17 and 0.18 mm in the case of lithotypes M1 and B1, respectively. Although these lithotypes represent different riverbed patterns, their riverbed was in each stage modeled under supercritical flow regime or in transitional circumstances towards such flow conditions (Fig. 10).

**Fig. 13** Types of channel deposits in relation to the flow velocity, channel bedforms and Weichselian fluvial lithotypes



ⓐ  meandering rivers and shallow braided rivers with upper plane bed (lithotypes M1 and B1)

ⓑ  braided rivers with unit and compound bars (lithotypes B2 and B3)

**I-III**  types of channel deposits

·······  limits of bedforms occurrence (after Ashley, 1990)

Unit bars and dunes, characteristic of the lithotype B2, are composed mainly of sands with a different gravel admixtures (Fig. 12). As a result, these sediments vary from fine- and medium-grained sands (characterized by symmetrical grain-size distributions) to coarse- and very coarse-grained sands, or even gravels (very positive and very negative skewed) (Fig. 12). The average value of median grain diameters is 0.45 mm.

The lithofacies of fluvial lithotype B3 comprise texturally homogeneous sands which contain of about 52% medium-grained sands. The average value of median grain diameters of these deposits is 0.26 mm (Fig. 12). These deposits are better sorted than sands of the lithotype B2, show symmetrical grain-size distributions and in few cases slightly positive or negative skewness. Sediments of the lithotype B3 have a negligible contents of coarse-grained sands and gravels and for these reasons, the average median grain-size diameter is lower than for the lithotype B2 and is 0.26 mm (Fig. 12). These data indicate greater stability of hydrodynamic conditions of the depositional environment of the lithotype B3 in comparison with the lithotype B2.

An analysis of the relationship between texture characteristics of the Weichselian fluvial successions in relation to the conditions for the development of the river bedforms, allowed for the delimitation of the three types of fluvial deposits. The first type ("I" in Fig. 13) is represented by the sand-bed-braided river with bar-related sedimentation ("b" in Fig. 13), transported mainly under condition of lower flow
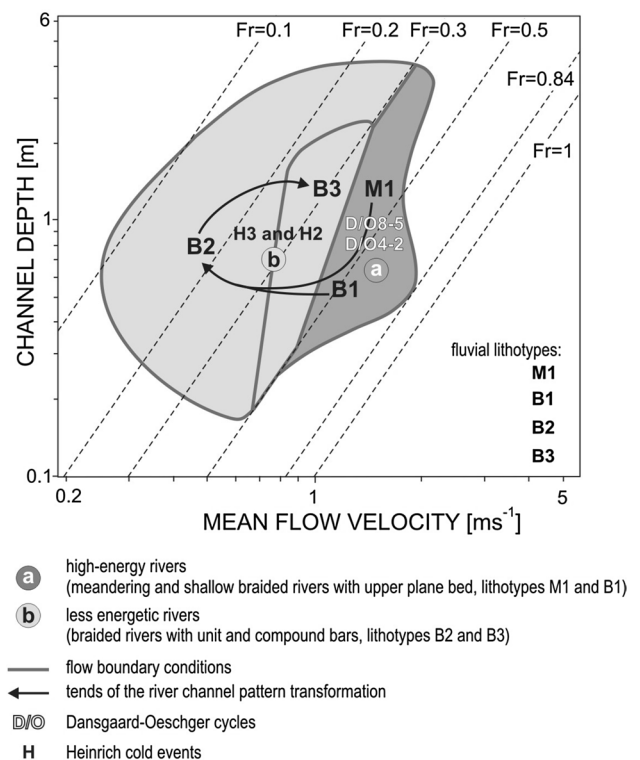


**Fig. 14** The model of river channel transformation during the Weichselian glaciation as an effect of climate oscillation

ⓐ  high-energy rivers (meandering and shallow braided rivers with upper plane bed, lithotypes M1 and B1)

ⓑ  less energetic rivers (braided rivers with unit and compound bars, lithotypes B2 and B3)

─── flow boundary conditions

◄── tends of the river channel pattern transformation

D/O  Dansgaard-Oeschger cycles

H  Heinrich cold events

regime (lithotypes B2 and B3). The sorting of these deposits becomes worse for sediments characterized by median grain diameter ($d_{50}$) in the range of 0.25–0.5 mm. The second type of the river channel deposits ("II" in Fig. 13) contains sands in which sorting is more poor with the decrease in median grain diameter from 0.25 to 0.15 mm. Moreover, these sediments represent gradual changes from the lithotypes related to sand-bed-braided rivers with unit and compound bars ("b" in Fig. 13) to the lithotypes of meandering rivers and shallow braided rivers with upper plane bed ("a" in Fig. 13), and hence changeable conditions from lower to the upper flow regime.

The third type of the channel sediments has an average grain-diameter ($d_{50}$) lower than 0.15 mm ("III" in Fig. 13). These sands represent high-energy, sand-bed and shallow-braided rivers (lithotype B1) or ephemeral meandering rivers (lithotype M1). The sorting of these deposits gets worse with the increase in size of median grain diameter.

As is apparent from the relationship between the median grain diameter ($d_{50}$) and the velocity of water flow ($V$), the same fraction of the deposits may be transported in different hydraulic conditions (see Ashley 1990; Southard and Boguchwal 1990; Southard 1991). This applies mainly to the deposits of an average diameter ($d_{50}$) smaller than 2.5 mm (types "II" and "III" in Fig. 13). Regardless of this, the median grain diameter of 0.15 mm may be the boundary between the channel sediments transported primarily in the conditions of the upper flow regime (type III of fluvial deposits) and deposits of sand-bed-braided rivers with transverse bars and dunes (types "I" and "II"; Fig. 13) (see Sly et al. 1983). Nevertheless, during the transportation of fine-grained sands ($d_{50}$ lower than 0.15 mm), there is a potential for ripples or an upper plane bed to develop (type III of channel deposits; Fig. 13), which refers flow competence to sediments transporting under conditions of different (lower or upper) flow regimes (see Ashley 1990; Southard and Boguchwal 1990; Southard 1991). For these reasons, the textual criterions ($d_{50} = 0.25$ and $d_{50} = 0.15$ mm) cannot be unambiguous in hydrodynamic interpretation of fluvial deposits.

## Climate-derived transformation of riverbed morphology and flow conditions during Weichselian glaciation

### Meandering and shallow-braided rivers

Lithotypes of the high-energy rivers, forming the lower members of the Weichselian fluvial formations, were deposited as a result of the activities of precipitation-dependent rivers (Weckwerth and Chabowski 2013). Typical water flow conditions of meandering and shallow-braided rivers (area

| AGE and H-EVENTS | | LITHOSTRATIGRAPHIC UNITS | LITHOTYPES | |
|---|---|---|---|---|
| Middle and Late Weichselian | H1 | Noteć Fm. | p3 | B2/B3/B4 |
| | | Starogród Fm. | | |
| | H2 | Zielonczyn Fm. | p2 | B3 |
| | | | | B2 |
| | | | | M1/B1 |
| | H3 | "Zielonczyn till" | | |
| | | Rzęczkowo Fm. (RFm) | p1 | B2 |
| | H4 | | | M1/B1 |
| | H5 | | | |
| | H6 | Chełmno Clays Fm. (CCFm) | | SS |
| Eeemian interglacial and Early Weichselian | | Lower Vistula Fm. (LPFm) | | B2? |
| | | | | M1/B1? |
| | | | | M |

**Fig. 15** The relation between lithostratigraphic units and Weichselian lithotype succession in the Toruń Basin [M—litothotype of eemian meandering river after Makowska (1980); SS—litothotype of proglacial ice-dammed lake after Wysota (2002); M1, B1–B3—fluvial litothotypes deposited during Weichselian glaciation (analyzed in this study); B4—fluvial litothotype of high-energy-braided river with sandy gravelly three-dimensional dunes after Weckwerth (2010, 2013); p1–p3—pradolinas 1–3]

M1/B1 in Fig. 14) were characterized by channel depth up to 1 m, a high Froude number (greater than 0.4) and high flow velocity (greater than $0.9 \text{ ms}^{-1}$). Apart from this, conditions of shallow water flow in meandering channels (lithotype M1) become similar to the conditions characteristic of shallow-braided rivers with the dominance of the upper plane bed (lithotype B1; cf. Dreyer 1993). Moreover, the increasing flow depth in meandering rivers, together with increasing their sinuosity index (SN from 1.25 up to 1.5), were responsible for decreasing the Froude number from 0.5–0.6 to 0.3–0.4 (Figs. 10, 11, 14). Increasing median grain diameter of the channel deposits in the case of the lithotype M1 (e.g., fluvial succession at the Chobielin site) reflects cumulative effects over successive cycles of

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climate changes (cf. Busschers et al. 2007; Van Balen et al. 2010), between Dansgaard-Oeschger cycles D/O8 and D/O5 (Rahmstorf 2003). In turn, this may have resulted in the seasonal increase in the channel aggradation rate of coarser sediments (Van Huissteden and Kasse 2001; Van Huissteden et al. 2001). Similar periodic floods are typical of modern rivers in the polar and sub-polar regions (Vandenberghe 2001).

Based on the thickness of the lithotypes M1 and B1 in relation to the time of their accumulation in the TSB, it can be assumed that the rate of channel deposition occurring between 28 and 24 ka ago (between D/O4 and D/O2) was higher than 40–30 ka ago (between D/O8 and D/O5). Its cause, despite the lack of meltwater feeding, may have been an increase in the efficiency of surface runoff affected by the expansion of continuous permafrost (Renssen and Vandenberghe 2003; Buschers et al. 2005, 2007). These climate conditions in the area of the Toruń Basin is proven by wedge casts filled with well-sorted sands at 27 ka ago (Drozdowski and Fedorowicz 1987; Drozdowski and Szupryczyński 1987).

Repetitive climate cooling at the beginning of H3 event (~30 ka ago) and H2 event (~24 ka ago) (Bond et al. 1999; Hemming 2004; Peck et al. 2007) caused changes in the paleoenvironmental conditions of river catchments. The hydraulic parameters of meandering rivers (lithotype M1) and shallow-braided rivers (lithotype B1) got similar to the conditions enabling threshold changes in the riverbed morphology (a.o. Schumm 1979). They consisted in decreasing the water flow velocity, the reduction of both hydraulic gradient and the Froude number, as well as in increasing the diameter of transported particles and the channel aggradation rate (Figs. 13, 14). The consequence was a river pattern transformation from meandering and shallow-braided- (lithotypes M1 and B1) to a braided river style with unit bars (lithotype B2; Figs. 14, 15). This process was initiated by humpback and two-dimensional dune development (e.g., Rozwarzyn site; Fig. 8).

**Braided rivers with unit and compound bars**

The lithotypes B2 and B3 form the upper members of the Weichselian fluvial formations. These lithotypes are related to the streams which were fed by meltwaters flowing from the north and precipitation-dependent rivers flowing from the south (Weckwerth and Chabowski 2013). Thus, lithotypes B2 and B3 represent two stages of the ice-marginal valley development (pradolinas 1 and 2; Fig. 15) during H3 and H2 events (pradolinas 1 and 2, respectively) which differ in relation to the lithological features.

The first stage began with the deposition of the lithotype B2. This corresponded to the transformations of the river channel pattern from meandering (lithotype M1) and

shallow-braided- (lithotype B1) into the sand-bed-braided river dominated by unit bars at the beginning of H3 and H2 events (cycles M1→B2 and B1→B2, respectively; Figs. 14, 15). These transformations were manifested by decreasing flow velocity (settled the low Froude number), linked to the energy losses associated with bedform roughness and with the transportation of coarser particles supplied to the Toruń Basin by sediment-laden meltwater from the north (Figs. 10, 11, 12, 13, 14).

The beginning of the second stage of the ice-marginal valley development (during H2 event, pradolina 2) coincided with unit bar amalgamation which resulted in the development of compound bars typical for the lithotype B3 (Figs. 14, 15). These macroforms constituted sand shoals within braided river channels. The water flows showed a similar depth range during the deposition of the lithotypes B2 and B3, although the flow velocity was generally greater in rivers of lithotype B3 (Figs. 11, 14), for which the energy of sedimentary environment was more stable than during the deposition of lithotype B2. Therefore, the lithotype B2 represents a transitional phase in the evolution of the sand-bed-braided river morphology characterized by two- and three-dimensional dunes and unit bar development.

Deposition of lithotypes B2 and B3 (which was related to the increase of the channel aggradation rate) took place 1–2 thousand years before every ice sheet advance into the Toruń Basin, during the development of pradolinas 1 and 2 (Figs. 14, 15). Fluvial lithotypes succession (B2→B3) was deposited during the middle Weichselian and indicates the relatively stable conditions of water flow only before the LGM (area B3 in Figs. 14, 15). Therefore, the development of compound bars is a mature and finale stage of the sand-bed-braided river evolution in fluvial cycle. Fluvial successions accumulated in the Toruń-Eberswalde ice-marginal valley (pradolina 3) during Pomeranian phase indicate braided river environment which varied under the terms of morphology and flow energy (Fig. 15; Weckwerth 2010, 2013; Pisarska-Jamroży 2015).

River channel pattern transformations which correspond to the fluvial lithotype sequences M1/B1→B2 and B2→B3, occurred 30 and 24 ka ago in the area of the Toruń Basin. Fluvial transformation proceeded in Central Poland and in the Southern North-Sea Basin at roughly the same time (Rotnicki 1987; Turkowska 1995; Huisink 1997, 2000; Mol 1997; Buschers et al. 2005, 2007; Cordier et al. 2006; Zieliński 2007). Their causes were related to the glacio-isostacy and climate cooling, which resulted in the development of continuous permafrost, in conditions of which increased the sediment supply into rivers in coincidence with particle coarseness in fluvial succession (cf. Huijzer and Vandenberghe 1998; Renssen and Vandenberghe 2003; Van Huissteden et al. 2001). Moreover, one of the main reasons for the transformation of the river channel morphology in the Toruń

Basin was river regime changes from precipitation-related to glacial which was associated with the two ice-marginal valley development before LGM (Weckwerth and Chabowski 2013). In these valleys, the high rate of fluvial accumulation (deposition of lithotypes B2 and B3) was dependent on the significant sediment load in meltwater streams which fed the Toruń Basin.

## Conclusions

The results of the sedimentological research of the Weichselian fluvial deposits in the Toruń Basin allowed the model of fluvial sedimentation to be recognized during successive phases of Weichselian ice sheet advances and retreats as a result of climate fluctuations. Sediments of sand-bed rivers are represented by four fluvial lithotypes which belong to two groups, the first of which (1) represents the high-energy meandering and shallow-braided rivers with upper plane bed (lithotypes M1 and B1—rivers of the precipitation-related regime) evolved during D/O8-5 and D/O4-2 cycles. The second group (2) corresponds to the braided rivers with unit and compound bars (lithotypes B2, B3—mainly rivers of the glacial regime) which existed at the time of the H3 and H2 events.

Stable conditions for the channel deposition were characteristic of rivers with different flow regimes and riverbed morphology. They were close to the upper flow regime for the lithotype M1 (ephemeral meandering river) and lithotype B1 (shallow braided and flat-bottomed river), while for the lithotype B3 (sand-bed braided river with compound bars)—to the conditions of the lower flow regime. Lithotype B2 represents the transitional phase in the evolution of the riverbed morphology, which corresponds to the sand-bed-braided river dominated by unit bars.

River channel pattern transformations are related to the changes of fluvial lithotypes (M1/B1→B2 and B2→B3) in Weichselian fluvial successions. It follows that the fluvial deposition that occurred as a result of the development of compound bars (lithotype B3), is a mature stage of the sand-bed-braided river development in the fluvial cycle. Transformations of the river channels during the Weichselian glaciation were caused mainly by the climatically related changes in the river regime and intensity of the surface runoff.

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