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MEANDER CORES ON THE FLOODPLAIN – THE EARLY HOLOCENE DEVELOPMENT OF THE LOW-FLOODPLAIN ALONG THE LOWER TISZA REGION, HUNGARY

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Abstract

The aim of the present study is to analyse the morphology, sedimentary structure and age of flood-free islands on the low-floodplain of the Lower Tisza in order to determine the date and causes of river incision.
On the study area the identified 16 elevated surfaces were divided into two groups. The real or meander core floodplain islands appear in the northern and central units of the study area. They are characterized by steep slopes, elevated surface and small territory (2.1 km^2 in average). The second type consists of elevated surfaces of point-bar systems and natural levees of paleo-channels, and they mostly appear in the southern unit. They have gentle slopes, smaller relative height and greater area (4.1 km^2 in average). The spatial distribution of the two types refers to slow and slight tectonic uplift in the northern part of the study area, though in the south they refer only to the lack of sinking.
The results of the sedimentological analysis on the meander core at Szegvár show that it originally belonged to the high-floodplain. Due to the slight tectonic uplift the meandering channel incised into the soft sediments, and as lateral erosion was possible, an ingrown-type meander developed, which later as a result of cut-off has become a meander core (or umlaufberg). Based on the OSL data the incision started at least 20.1±2.1 ka ago, and it terminated ca. 8-9 ka ago. The calculated bankfull discharge of the Szegvár paleo-meander is estimated to be 4000-7500 m^3/s, referring to a considerably higher discharge than that of the present-day Tisza (800 m^3/s). Similar planimetric meander parameters of paleo-channels on the high and low-floodplain suggest that the incision was not driven by climate, i.e. discharge change but primarily by tectonic movement. This is also supported by the height condition of the islands, as their surface is almost at the level of the high-floodplain.

Keywords: floodplain formation, incision, ingrown meander, floodplain island, OSL dating

INTRODUCTION

The floodplain of the Tisza River south of Csongrád (Hungary) was formed at the Late Pleistocene and Holocene through the alteration of erosional and aggradational activities along the river. The most important event in its development history was an incision, which resulted in the differentiation of the low and high-floodplain. According to Láng (1960) and Mezősi (1983) this can be dated to the beginning of the Holocene, approximately 12 ka ago, when the Tisza and Maros Rivers incised into the former floodplain by ca. 10-12 m. They explained the process by an increased erosional activity, however the causes of intensified fluvial activity were not resolved. According to Mátyus (1968) the incision occurred later, in the Boreal Phase and it resulted in the development of a 3-5 km wide floodplain. However, Lovász (2002) contested this idea as he claimed that the amount of precipitation had decreased in the Boreal Phase, consequently run-off and fluvial processes became limited. Based on the paleo-discharge data of Gábris (1985, 1986) the discharge of the rivers was mush higher (up to 10 times) in the Preboreal and Atlantic Phases, which could also explain the intensive incision. According to Gábris (1995), the development of the low-floodplain is a rather new process (ca. 4000-5000 y), and overbank accumulation had also started recently, ca. 2.5 ka ago.

The explanation of the incision and subsequent aggradation just by climatic causes has several week points. The area is quite active tectonically: the area of Szeged is the most intensively subsiding area of the great Hungarian Plain since the beginning of the Neogene (Borsy, 1990). Based on the lack of overbank aggradation, however, Somogyi (1967) supposed the termination of sinking since the end of the Pleistocene, or even since the beginning of the Riss-Würm Interglacial. Nevertheless, present-day measurements (Joó, 1998) on vertical tectonic movements indicate intensive subsidence (3-4 mm/y) in the southern part of the Lower Tisza Region, especially at the city of Szeged, at the confluence of the Tisza and Maros Rivers. Another centre of the sinking is at the Middle Tisza Region, where the rate of subsidence is similar to that experienced along the Lower Tisza. In between the two sinking areas, in the northern part of the Lower Tisza Region tectonic movements have a lower rate, and thus the area is relatively up-lifted. However, Timár (2003) emphasizes, that the intensive exploitation of subsurface waters, natural gas and oil might also cause surface subsidence, thus the present-day rate of tectonic movements should not projected to longer periods.

Consequently, no precise data exist on Quaternary vertical movements, but the tectonic activity is clearly indicated by the fluvial forms on the low-floodplain. In case of subsidence overbank aggradation becomes dominant, which can also hide previous forms (Twidale, 1964). In case of uplifting fluvial erosion intensifies, causing incision and subsequent lateral erosion on the former floodplain. Hence, meander cores can develop on the low-floodplain as the result of (ingrown) incised meander development (Fairbridge, 1968), and the process can also separate the low and high floodplains (Gábris, 1995). Incised meanders have two types: ingrown and entrenched (Morisawa, 1985), both developing as a result of tectonic uplift (Twidale, 1964). In case of entrenched meanders incision is quick and meander development usually reaches the solid rock bed, while ingrown meanders develop usually at a slower incision rate when lateral and vertical erosion can coexist. As the result of lateral erosion the meanders can be cut off, which results in the development of meander cores being much higher than the alluvium. Meander cores are cited in different ways in the literature: pembina (Fairbridge, 1968), bedrock spur or natural arch (Summerfield, 1991; Hugett, 2007), umlaufberg (Mahaney, 1984). As incision and lateral erosion coexist, the cross-section of the valley is asymmetric: the convex side of the meander has gentler slope and on the slip-off face indications of point-bars might appear (Fairbridge, 1968).

Meander cores of ingrown meanders appear on the low-floodplain of the Lower Tisza Region. Until the 19th century river regulation works they were flood-free surfaces surrounded by marshlands, therefore in earlier researches these elevated surfaces were cited as floodplain islands, flood-free islands, terrace-islands or Pleistocene remnant surfaces (Mátyus, 1968; Andó, 1969; Mezősi, 1983). The meander cores can be the key to the reconstruction of the first stages of the development of the Lower Tisza Region. The aim of the present study is to identify and classify the flood-free islands of the low-floodplain, to determinate the date of incision and its causes in the Lower Tisza Region by studying the morphology, sedimentary structure and OSL age of meander cores.

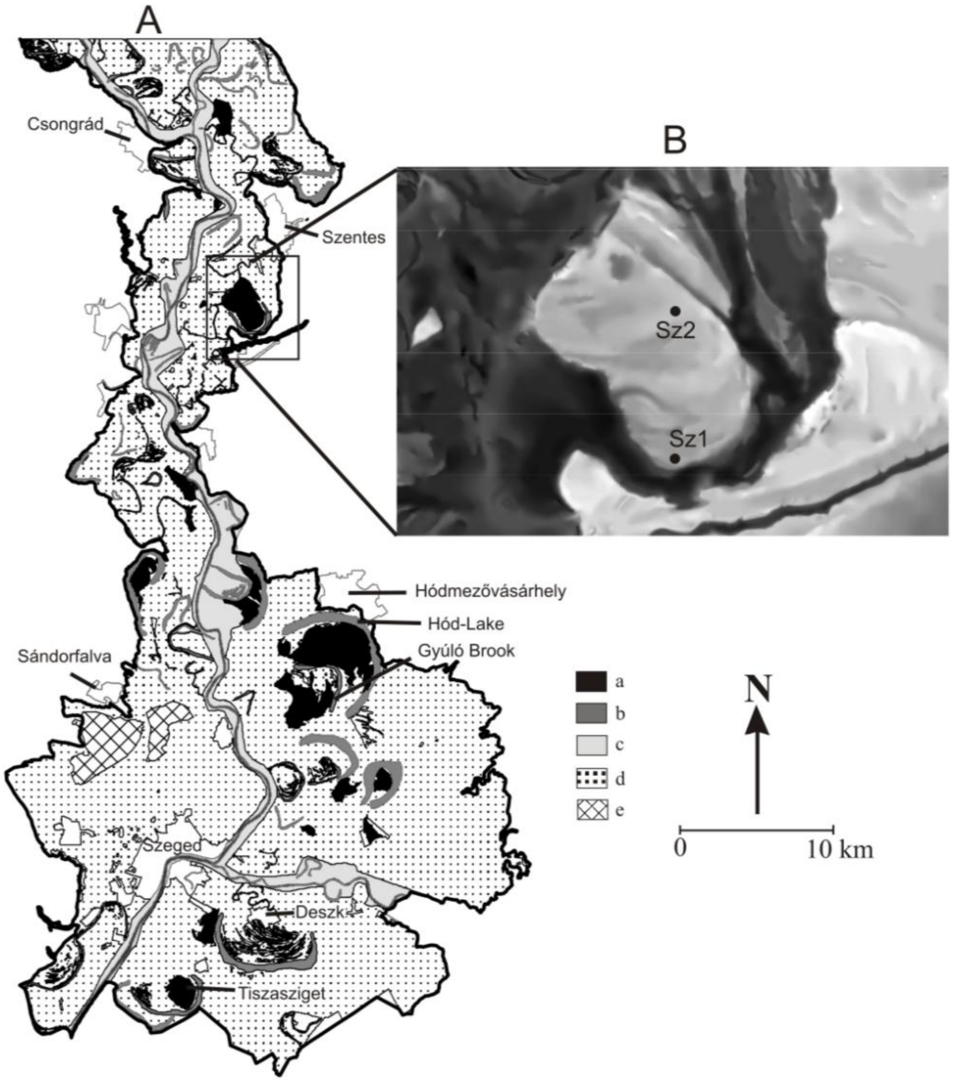

Fig. 1 Main geomorphic features of the Lower Tisza Region (A) the meander cores near Szegvár were studied in detail (B) a) elevated surface, b) paleo-channel, c) artificial floodplain, d) alluvium, e) lake, SZ1-2 sampling sites

STUDY AREA

The research was carried out on the low-floodplain of the Lower Tisza Region (territory: 1214 km^2) between Csongrád and the Hungarian state border (*Fig. 1A*). The border between the low and high-floodplain is quite pronounced (1-5 m) on the northern part of the study area (north of the line between Hódmezővásárhely and Dóc), though on the southern part it is less striking (0.5-2 m) (Kiss and Hernesz, 2011).

Most of the low-floodplain was annually flooded until the 19th century river regulation works (Kiss et al., 2011) therefore it is covered by young overbank sediments (Marosi and Somogyi, 1990). Based on the geomorphic features of the low-floodplain, the study area could be divided into three units (Kiss and Hernesz, 2011). The northern unit (north of the Csongrád-Szentes line) has an elevated edge (3-5 m), it gets narrower towards the south (average width: 13.3 km) and paleo-channels and point-bar systems cover its surface. The central unit (north of the Hódmezővásárhely-Sándorfalva line) is the narrowest (average width: 8.1 km), though its edge is quite sharp (3-5 m). In the central unit scour channels (breaches) and back-swamps are the dominant fluvial forms, indicating limited lateral migration. The southern unit is the widest (average width 24.8 km) and least marked (1-1,5 m in the east and 1-3 m in the west). Its eastern part is covered by paleo-channels and well-developed point-bar systems, however its rim is not well defined as the alluvial fan of the Maros stretches into it. The western part of the southern unit is poor in fluvial forms, as its surface is covered by 3-5 m thick Pleistocene loess (Miháltz, 1967; Mátyus, 1968). Therefore, mostly based on bio-geographical evidences Deák (2010) excluded this part of the low-floodplain. In this case the average width of the southern unit is only 10-15 km, and its western edge is only 1-2.5 m high.

METHODS

A digital terrain model (DTM) was created based on topographical maps (scale 1:10.000), using ArcGIS 10. On the DTM the floodplain islands were identified, their rims and forms, and the neighbouring paleo-channels were localised. The morphomertric parameters of the islands were calculated (area, maximum width and length, mean relative height over the low-floodplain and mean height difference from the closest high-floodplain), and also the mean height of the low and high-floodplain in the vicinity of the floodplain islands were determined. The radius of curvature, width and meander length of the paleo-channels embracing the islands were also measured, and based on these horizontal parameters the paleo-discharge of the channels were determined using the equations of Timár and Gábris (2008).

On the meander core near Szegvár (*Fig. 1B*) two corings (400 cm deep) were made. The SZ1 coring was established on the bank of the paleo-channel, the SZ2 coring was made on the slip-off surface of the island, on the remnants of its former point-bar (*Fig. 5a*). The grain-size distribution of the samples was determined by Fritsch Analysette 22 laser equipment with a measurement range of 0.08-2000 μm. Samples underwent ultrasonic homogenisation and all measurements were repeated three times to check if there is further disintegration. Grain-size classes were determined using the Udden scale.

The age of samples was determined by optically stimulated luminescence (OSL). From the SZ1 and SZ2 corings four and three OSL samples were collected, respectively. Due to the inadequate amount of medium sand in the samples fine grains (4-11 μm) were separated and dated. Although fluvial sediments are examined usually by using the coarse grain method (Rittenour, 2008), we found earlier that in a medium energy environment fine grain mean and coarse grain minimum ages are fairly well corresponding (Sipos et al., 2010) in the area. Nevertheless, based on the measurement of fresh, known age point bar sediments along the Tisza River, ages derived from silty deposits may overestimate the true age by 0.5-1.0 ka (Kovács, 2011), which has to be considered.

The carbonate and organic content of the samples were removed by HCl and H_2O_2. Finally samples were treated in hexafluoro-silicic acid to remove feldspars and to receive the quartz fraction of the sample. Samples were settled on aluminium discs from an acetone suspension. Measurements were made on a RISØ TL/OSL DA-15 automated luminescence reader using the Single Aliquot Regeneration (SAR) protocol (Wintle and Murray, 2006). Presence of residual feldspar was monitored using the OSL/IR depletion ratio (Duller, 2003). Pre-heat tests were carried out in the 190-280 °C range, all samples yielded 220 °C as the suitable preheat temperature for the OSL measurements. In situ gamma and cosmic dose rate was measured by a Canberra Inspector 1000 portable NaI gamma spectrometer during sampling. Alpha and beta dose rates were determined on the basis of U, Th and K contents measured by a Canberra-type laboratory gamma spectrometer equipped with a low background coaxial Ge detector. Alpha efficiency was taken 0.1, wet dose rates were calculated on the basis of in situ water contents.

RESULTS AND DISCUSSION

Morphometry of the floodplain islands

On the study area altogether 16 elevated surfaces were identified on the low-floodplain. These floodplain islands were divided into two groups based on their morphology and origin: (1) meander cores and (2) the high point-bars and natural levees of the paleo-meanders.

The first group of islands have sharp rims and were eroded on some sides by paleo-channels. The mean territory of these islands is 2.1 km^2. The largest (6.3 km^2) is located near Szegvár (No.3), and was studied in detail, whilst the smallest (0.1 km^2) is in its close vicinity (No.2). These islands are slightly elongated, as their average length/width ratio is 2.4. On their slip-off slope point-bars remnants of could be identified. Their edge is well-defined (*Fig. 2*), since they are laterally eroded by subsequent paleo-channels. The size of these paleo-channels (mean Rb= 996 m) is much greater than that of the present-day Tisza (mean Rb= 600 m).

The average height of meander cores is 81.3 m asl, the highest (83.6 m asl) is the northernmost one near Magyartés (No.1), and the lowest (79.2 m asl) can be found in the southern unit, near Maroslele (No.13). The relative height of the islands compared to the low-floodplain is 2-4 m. The highest ones (No.1 and 15) has a relative height of 4.1 and 4.0 m, though the lowest (No.13) is also 2.2 m above the low-floodplain level. Their height reaches or in some cases exceeds the height of the neighbouring high-floodplain. For example the island at Szőreg (No.15) is inhabited since the Neolithic period (since ca. 6500 BP) as the flood-free are provided good circumstances for settlement, and also a 12-13th c. monastery was built on it (Trogmayer, 1977). Thus the surface of the meander cores could be further elevated as a consequence of human activity.

These meander core islands appear mostly in the northern and central units of the Lower Tisza Region, and only two of them (No.15 and 16.) are located in the southern unit.

The second type of floodplain islands consist of elevated point-bars and natural levees of the paleo-channels, thus their rims are not well-defined. The average area of these islands is 4.1 km^2, the largest (16.2 km^2) is the extensive point-bar system of the Hód-tó paleo-channel (No.9), and the smallest (0.4 km^2) is island No.14. Their form is elongated, the greatest length/width ratio is 4.5 (No.7), which is connection with their fluvial origin. On their surface 1-2 m height differences appear, showing the location of point-bar ridges and swales (*Fig. 3*).

The highest (80.6 m asl) of them is near Dóc (No.6), while the smallest (78.9 m asl) is near the Gyuló paleo-channel (No.10). Their height compared to the level of the low-floodplain is smaller than of the first island group, as the highest (No.6) and lowest (No.8) representatives are 3.2 m and 1.3 m high, respectively. Their surface is much lower (79-80.5 m) than the level

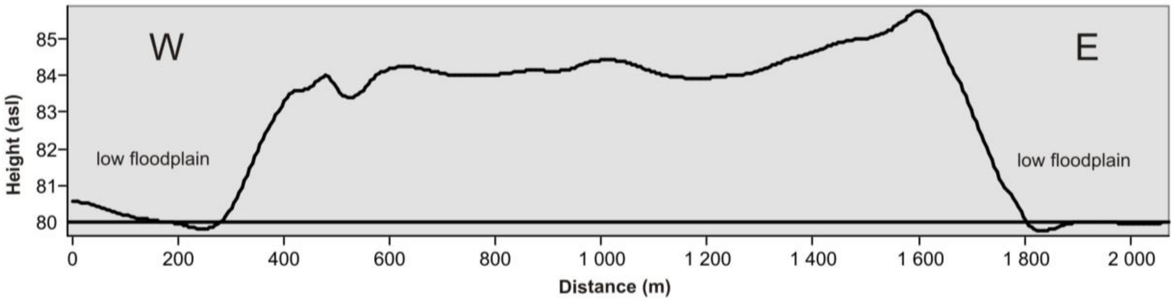

Fig. 2 The cross-sectional profile of island No.1 showing the eroded rims and slip-off face with the remnants of point-bars

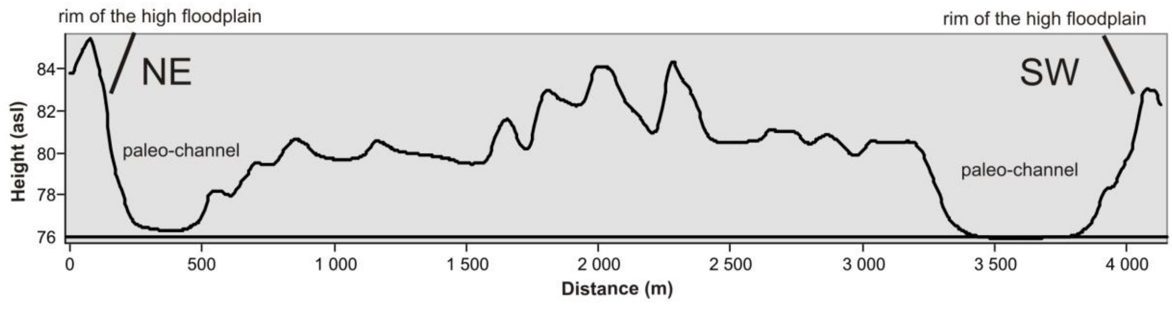

Fig. 3 The cross-section of island No.6, representing an elevated (2-4 m) point-bar system, where the brink-line of the point-bars is almost as high as the high-floodplain of the high-floodplain. They appear along those paleo-channels (e.g. Hód-tó, Gyúló-ér, Deszk, Tiszasziget), which had quite high bankfull discharges. Their radius of curvature (570-3200 m) is 2.5 times greater than of the present-day Tisza (Fiala and Kiss, 2006). They mostly appear in the southern unit of the study area, where these paleo-channels remained

The two island types have characteristic spatial distribution (*Fig. 4A*). In the northern and central units of the study area islands are much higher than the low-floodplain and their surface is almost at the same level as of the high-floodplain. Their steep rims are well defined. Their similar height to the high-floodplain indicates that they can be defined as meander cores, and they are relatively young features as they were not eroded (Andó, 1969). In contrary, the point-bar and natural levee islands appear along the paleo-channels in the southern unit of the study area. These elevated surfaces have smaller relative height and gentle slopes.

The spatial distribution of these two island types explains some details of the development history of the region. The meander cores refer to slow incision, whilst the point-bar and natural levee islands indicate lateral floodplain aggradation, where the subsequent overbank accumulation terminated or became very slow, as the forms are not buried by fluvial deposits, and the surface is not evened out.

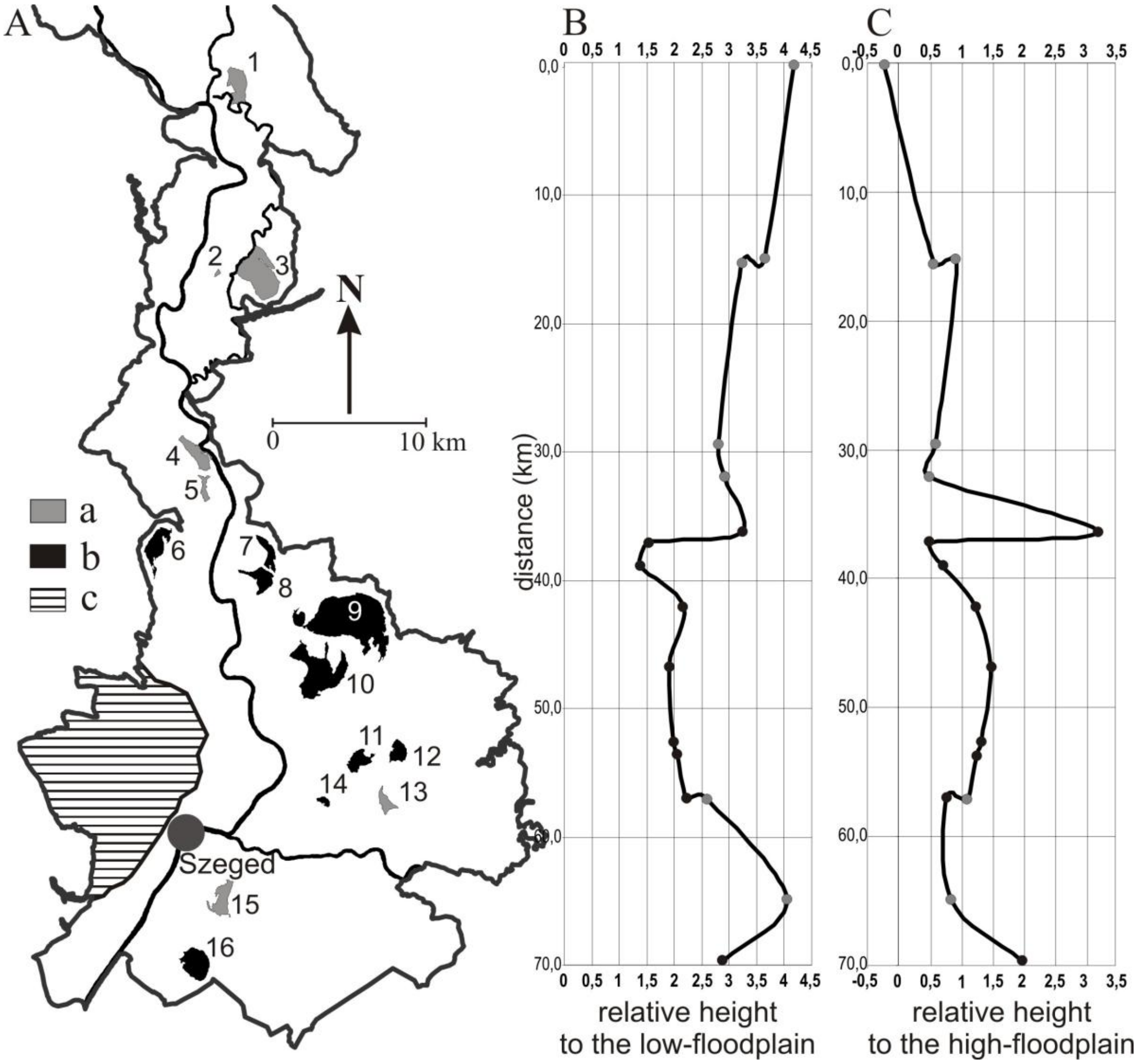

Fig. 4 Location and type of floodplain islands and their relative height

A) Location of floodplain islands: a) meander cores, b) elevated point-bars and natural levees of paleo-channels, c) Pleistocene loess blanket, No.1: Magyartés; No.2: Tétel-hát, No.3: Szegvár, No.4: Ányás I., No.5: Ányás II., No.6: Dóc, No.7: Körtvélyes, No.8: Solt, No.9: Hód-tó, No.10: Gyúló-ér, No.11: Nagyfa, No.12: Batida, No.13: Maroslele, No.14: Lebő-halom, No.15: Szőreg, No.16: Tiszasziget; B) Relative height of islands compared to the low-floodplain.; C) and to the high-floodplain

The spatial distribution of the height condition of floodplain islands

The relative height of island surfaces compared to the high and low floodplains reflects the tectonic activity during and after their development. The height conditions of the western and eastern high-floodplain surfaces are not similar, thus the height of the floodplain islands was compared to the closest high-floodplain surface.

The relative height of the floodplain islands (calculated from the level of the lower floodplain) decreases towards the south (*Fig. 4B*), though this tendency is broken by some islands, as islands No.6 and No.15 are slightly higher and No.7 and 8 are lower. In case of island No.6 this can be explained by the surrounding deeper-lying paleo-channels, resulting a relatively greater height value, whilst in case of island No.15 (at Szőreg) this can be explained by anthropogenic activity (Trogmayer, 1977). The smallest height difference between the lower floodplain and floodplain islands was found in case of the islands No.7 and 8 (at Körtvélyes and Dóc), which are higher than the lower floodplain only by 1.3 and 1.5 m, respectively.

The highest islands (over 3.0 m) mostly appear in the northern and central units of the study area, whilst the smallest ones (under 2.5 m) are located in the southern unit, south of the line between Hódmezővásárhely and Dóc.

The height of islands compared to the closest high-floodplain surface refers to their evolution: supposing that the present-day high-floodplain was the former floodplain of the Tisza (Mezősi, 1983), the surface of islands represents the same level. These forms were formed during the incision of the river into the high-floodplain and simultaneous development of the low-floodplain The mean height of the islands was extracted from the mean height of the high-floodplain (*Fig. 4C*). Only one negative data was measured (No.1, Magyartés): here the island is higher than the high-floodplain by 0.25 m. Towards the south the islands No.1-8 are almost as high as the high-floodplain (they are lower only by 0.5-0.9 m). The next island (No.6) is unique, considering its height conditions, as it is much higher than the low-floodplain (3.2 m) despite of its great difference (3.1 m) from the neighbouring high-floodplain (reason I explained earlier).

Going south the height difference between the high-floodplain and islands surfaces increases (1.0-1.5 m) which can be explained by the different origin of the islands (point-bar systems and natural levees). Island No.13 (at Maroslele) and island No.15. (Szőreg) are the only exceptions being only 0.7-0.8 m lower than the high-floodplain.

Based on relative height conditions, islands in the northern and central parts are much higher, and are at the same level as the high-floodplain. Meander cores in the northern unit imply an uplift process resulting incision and the development of ingrown meanders. On the contrary, islands in the southern part of the study area are less elevated, and the southern unit is a low lying alluvium, however, sinking could not be characteristic or could not last long, as overbank sedimentation did not cover the floodplain forms. All these suggest that the southern part of the study area developed in a different way than the northern and central parts.

Detailed analysis of the No.3 meander core near Szegvár

The area of this meander core is 6.3 km^2, its greatest width is 2.0 km and greatest length is 3.2 km (*Fig 5a*), its mean height is 81.9 m asl, its highest point is at 85 m asl. Its north-eastern rim is quite steep, it was laterally eroded by the Kurca Brook. On the other sides of the form rim is well defined, though it has more gentle slopes. The surface of the studied meander core is not uniform, as the rims are dissected by gullies and on its slip-off surface the fragments of point-bars left behind by incising paleo-meander can be identified (*Fig. 5b*). The paleo-channel dividing the high-floodplain and the surface of the meander core is wide (300-500 m), the length of the palaeo-meander is ca. 5-7 km. Based on its morphometric parameters, paleo-discharge (4000-7500 m^3/s) was much higher than that of the present-day Tisza (800 m^3/s). In this paleo-meander a smaller misfit channel of the Kurca Brook can be identified.

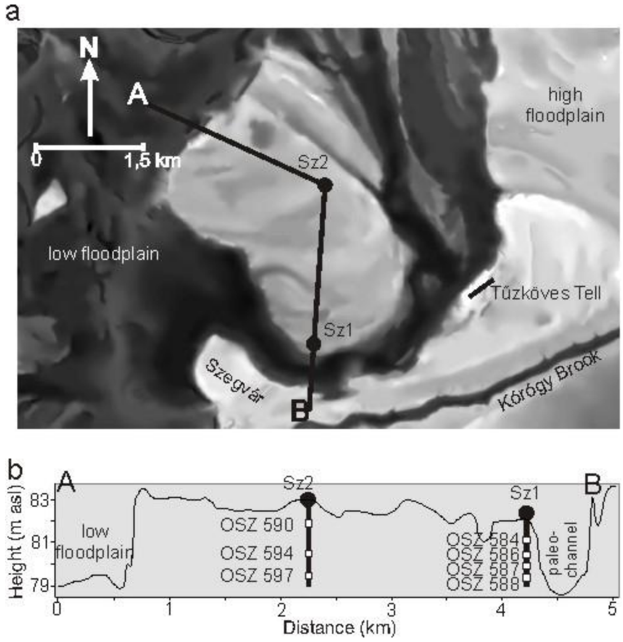

Fig 5 The digital elevation model of floodplain island No.3 at Szegvár, the location of corings (a) and the cross-section of the meander core (b)

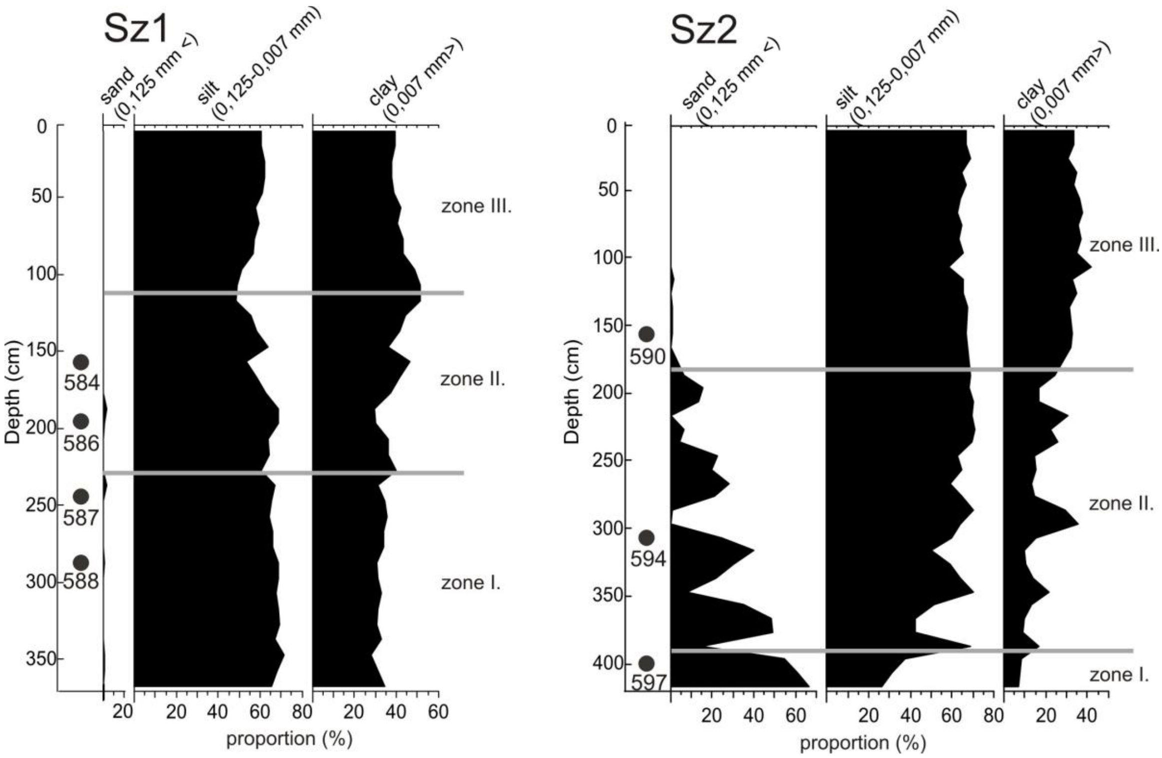

Fig. 6 Grain size distribution of the two corings (Sz1 and Sz2) and the depth of OSL samples

The neighbouring high-floodplain (82.9 m asl) is ca. 4 m higher than the low-floodplain (78.7 m asl), the arched rim in between is well defined as it was laterally eroded by the paleo-channel. The surface of the high-floodplain is dissected by the headward erosion of some channels (i.e. Kórógy Brook).

The high-floodplain probably has not been reached by floods as early as the beginning of the Holocene, thus in the Neolithic Period a continuous settlement could be established on the top of the rim, as it is indicated by the Neolith tell of Szegvár-Tűzköves dated back to 6.8-7.1 ka (Korek, 1987). But the paleo-channel beside the tell and around the meander core remained a swampy area until the late 19th c. river regulation works (on old maps it was indicated as Kontra Lake).

Two corings were made in order to analyze the meander core from a sedimentological aspect. The Sz1 coring was made on the bank of the paleo-channel, approx. 100 m far from the rim of the meander core. In the whole profile clay and silt are dominant. In the lower zone (230-370 cm) clay content (ca. 60 %) overwhelms the silt content (ca. 30 %) (*Fig. 6*). In the middle zone (110-230 cm) the clay content increases further with three well-defined peaks. In the uppermost zone (0-110 cm) the silt becomes dominant (over 60 %). Fine sand (under 0.06-0.125 mm) appear only occasionally (at 180-190 and 230-240 cm), but its proportion is never over 2 %. Results suggest, that these sediments were deposited at low energy conditions: probably the profile represents the loessy overbank deposits (or channel infill) of the paleo-channel.

From Sz1 zone I. and II. four OSL samples were taken (*Table 1*). The oldest sample (190-200 cm) is 16.9±1.7 ka, while the youngest (150-160 cm) is 15.7±1.5 ka old. Thus, the overbank or channel-infill sediments of the lower part of the profile were deposited around 15-17 ka ago at the end of the Ságvár-Lascaux Interstadial or at the beginning of the Oldest Dryas (The geological times scale compiled by Gábris (2003) was applied).

The Sz2 coring was made on the slip-off face of the meander core on the top of a remnant point-bar. In the lower zone of the profile (390-420 cm) sand fraction is dominant (55-60 %), the proportion of silt (26-37 %) and clay (6-7 %) is smaller. In the middle zone (180-390 cm) four peaks of the sand fraction appear (at the depth of 350-380, 300-340, 240-280 and 180-210 cm), but the proportion of sand in the peaks is continuously decreasing from 48.8 % to 15.5 %. At the same time the amount of silt is increasing (from 26.4 % to 70.2 %). In the upper zone (0-180 cm) the sand almost disappears and clay becomes dominant (over 60 %) overtaking silt (ca. 30 %), similarly to coring Sz2. In the lower and middle zones the sand peaks refer to

Table 1 Dosimetry, equivalent dose and OSL age data of the investigated silt samples

site	sample	depth (cm)	water content (%)	U-nat (ppm)	^{232}Th (ppm)	K (%)	$D'_{\alpha+\beta}$ (Gy/ka)	$D'_{in\ situ}$ (Gy/ka)	D'_{total} (Gy/ka)	D_e (Gy)	OSL age (ka)
Sz1	OSZ584	150-160	18 ± 1.8	1.79	6.78	1.31	1.83±0.24	0.89±0.04	2.00±0.24	36.99±1.44	**15.7±1.5**
	OSZ586	190-200	24 ± 2.4	1.64	7.28	1.36	1.74±0.22	0.89±0.04	1.91±0.23	38.57±2.06	**16.9±1.7**
	OSZ587	240-250	24 ± 2.4	1.94	7.25	1.49	1.90±0.25	0.78±0.03	2.07±0.25	37.66±2.10	**16.2±1.8**
	OSZ588	280-290	25 ± 2.5	1.91	7.67	1.42	1.86±0.24	0.78±0.03	2.03±0.24	36.61±1.90	**16.0±1.7**
Sz2	OSZ590	150-160	13 ± 1.3	1.72	6.1	1.07	1.70±0.22	0.84±0.04	1.87±0.22	31.73±0.83	**14.4±1.3**
	OSZ594	310-320	15 ± 1.5	1.55	5.75	0.92	1.49±0.19	0.74±0.03	1.66±0.19	35.13±3.37	**18.1±2.4**
	OSZ597	400-410	17 ± 1.7	1.2	4.2	0.78	1.15±0.15	0.57±0.02	1.32±0.15	30.21±2.36	**20.1±2.4**

near-bank deposits, which were deposited during high energy conditions. These were covered by suspended sediments at falling stage or at smaller floods. The samples of the upper zone represent overbank floodplain sediments.

OSL dating was carried out on three samples of coring Sz2. The lowest sample (400-410 cm) probably represents the active point-bar, the sandy samples at 310-320 cm stand for near-bank deposits, whilst the fine grain dominated zone at 150-160 cm correspond to overbank floodplain sediments (*Fig. 6, Table 1*). The age of the active point-bar, 20.1±2.1 ka, dates back to the Late Glacial Maximum. The accumulation of the near-bank deposits took place at 18.1±2.1 ka, in the Ságvár-Lascaux Interstadial. The clayey-silty overbank sediments accumulated 14.4±1.3 ka ago, thus it was deposited simultaneously with the youngest samples of coring Sz1 – considering their error intervals.

Based on the morphological, sedimentological and OSL data of meander core No.3 at Szegvár the incision of the paleo-meander was already in progress in the Late Pleniglacial. The river laterally eroded the present-day edge the high-floodplain slowly and deposited its sandy point-bar on the slip-off surface of the meander core. This result fits to the findings of Domokos and Krolopp (1997), who made a sedimentological and paleontological study on a sandy deposit on the high-floodplain. They found fluvial sandy deposits covered by 16-18 ka old loess. This reinforces our data, as the point-bar and near-bank sandy layers of the studied paleo-channel are 18.1-20.1 ka old, thus this meander probably indicates the very beginning of the incision process. The over-bank (or channel infill) loessy-silty sediments of the corings date back to 14.4±1.3 and 16.9±1.7 ka, thus by this time incision was significant, and only overbank sediments were deposited on the surface of the meander core. The age of these sediments correlates well to the results of Krolopp et al. (1995), studying the loess blanket west of Szeged, and finding that the loessy cover over the fluvial sediments was deposited ca. 13-25 ka ago.

CONCLUSIONS

Two types of floodplain islands were identified in the Lower Tisza Region. The real or meander core floodplain islands are characterized by steep slopes, elevated surface (higher than 80.5 m) and small territory (2.1 km^2 in average), and they appear in the northern and central units of the study area. They were identified as meander cores of earlier paleo-channels. The second type consists of elevated surfaces related to point-bar systems and natural levees of paleo-channels. They are characterized by gentle slopes, smaller relative height (under 2.5 m) compared to the low-floodplain and greater area (4.1 km^2 in average), and they mostly appear in the southern unit.

The spatial distribution of the two types of floodplain islands can be explained by the different evolution of the floodplain. The development of meander cores in the northern and central units refer to slow and slight tectonic uplift, though in the southern unit the existence of higher fluvial forms of old paleo-channels refer to the lack of sinking, otherwise they would be covered by overbank sediments. This result is in contradiction with the present-day rate of sinking (Joó, 1998), but this disagreement could be explained by the intensive subsurface natural oil, gas and thermal water extraction causing subsidence (Timár, 2003).

The results at the meander core at Szegvár are corresponding well to the data of earlier sedimentological analyses carried out on the high-floodplain (Domokos and Krolopp1997), and prove that the meanader cores are remnant surfaces of the high-floodplain. The age (14.4±1.3 ka to 16.9±1.7 ka) of the loessy overbank sediments covering the meander core date back to the Oldest Dryas, when fluvial loess deposited intensively in the area (Sümegi and Krolopp, 1995; Szöőr et al., 1991).

Due to the slight tectonic uplift the meandering channel incised into the soft sediments, and as lateral erosion was possible, an ingrown meander developed, which turned to be a meander core following a cut-off event. The OSL age of the uppermost point-bar of the ingrown paleo-channel indicates that the incision started at least 20.1±2.1 ka ago. However, the last activity phase of a paleo-channel (at Kenyere Brook) on the high-floodplain was dated back to only 11.5±0.8 ka (Sipos et al., 2010). The incision terminated at 8-9 ka ago, as a huge paleo-meander near Deszk remained well preserved on the low-floodplain, and it was forming its youngest point bars at 8.9±0.6 ka ago (Sipos et al., 2010).

The incision was probably a rapid process. It lasted only for 1-2 ka, as the radius of curvature of paleo-channels on the high-floodplain and at the feet of the meander cores is similar. The rapid incision is also supported by the similar OSL ages of point-bars of the paleo-channels on the high- and low-floodplain (Sipos et al., 2010).

The calculated bankfull discharge of the paleo-meander at Szegvár was 4000-7500 m^3/s, referring to a considerably higher discharge than that of the present-day Tisza (800 m^3/s). This discharge is similar to the discharge of the paleo-meander on the low-floodplain near Deszk (7000 m^3/s) but lower than the paleo-channel (12.700 m^3/s) along the Kenyere Brook on the high-floodplain (Sipos et al., 2010). The similarity in the horizontal morphometric meander parameters of paleo-channels on the high and low-floodplain suggest that the incision was not driven by climate change but primarily by tectonic movement. It is also supported by the height condition of the meander cores, as their surface is almost at the level of the high-floodplain.

Consequently, the northern and central units of the study area were characterized by slight tectonic uplift in the Late Glacial or early Holocene periods, resulting slow incision. In contrary the southern unit was lower lying, less active tectonically, and subsidence was probably not characteristic until nowadays.

Acknowledgements

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MORPHOLOGICAL AND HYDROLOGICAL CHARACTERISTICS OF PALEO-CHANNELS ON THE ALLUVIAL FAN OF THE MAROS RIVER, HUNGARY

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Abstract

The aim of our research was to identify and map the paleo-channel systems on the alluvial fan of the Maros River and to analyse their spatial characteristics. The study on flow directions, horizontal channel parameters and paleo-discharge of the channels can help to forecast the maximum flood discharge and channel changes influenced by climate variations. The paleo-channel generations on the Maros alluvial fan form 13 zones with well defined boundaries. These zones can be either dominated by meandering (5), braided (2), or the mixture of meandering and braided patterns (3). The remaining three paleo-channel zones exhibit an anastomosing pattern but they were not analysed in this study. The horizontal morphological parameters of the braided, the meandering and the misfit channels were measured. Based on these morphometric parameters and regional discharge equations the bankfull discharge of the meandering zones was calculated. The greatest discharge was around 2655 m^3/s while the smallest was 27 m^3/s in case of a misfit paleo-channel. Based on the slope conditions the alluvial fan was divided into three parts. The greatest slope (31.0 cm/km) was found in the central part of the alluvial fan, whilst slightly lower slopes (23.8 cm/km and 24.9 cm/km) characterise its axial and distal parts. These parameters refer to a normal radial profile of an alluvial fan. The channel pattern changes are in close relation with differences in slope. This is the most obvious in zone No. IX, where braided channels transform into meandering and then braided again from east to west in accordance with slope conditions.

Keywords: paleo-channel, paleo-discharge, morphometric parameters, slope conditions

INTRODUCTION

Paleo-hydrological data play an important role in understanding the response of rivers to climate change. For example precipitation and runoff conditions can be reconstructed by calculating paleo-discharge data. There are several methods to reconstruct the hydrological conditions of the past. Some studies use proxy data, for example the ratio of stable oxygen isotopes or the rate of the high-resolution magnetic susceptibility, but morphometric parameters of alluvial channels can provide more precise data (Stein et al., 2004; Carson and Munroe 2005; Scheurle et al., 2005; Saenger et al., 2006).

Scheurle et al. (2005) applied paleo-oceanographic stable oxygen isotope ($\delta^{18}O$) to determine paleo-discharge of the Elbe River. They analysed the rate of isotopes in calcareous shells of marine animals and correlated it to salinity, thus the ratio of salty- and freshwater (referring to paleo-discharge) were estimated. Saenger et al., (2006) applied the same method, but besides they modelled the precipitation conditions of the drainage area too.

Sedimentological, geochemical and micro-paleontological proxy data of surface sediments also allow the reconstruction of paleo-climatic conditions. High-resolution magnetic susceptibility record was used to estimate sediment fluxes and their relationship to paleo-environmental changes by Stein et al. (2004). The variability of sediment fluxes during the Holocene was related to the changes in discharge and coastal erosion input.

Carson and Munroe (2005) applied dendro-hydrology to reconstruct mean annual discharge and precipitation data. The width of an annual tree ring reflects the yearly hydrological conditions and also refers to the near-surface temperature, evapo-transpiration and precipitation (Werritty and Leys, 2001).

Since there is a close relationship between discharge and different channel parameters, a wide range of paleo-discharge calculations exist (Gábris, 1970, 1986; Sylvia and Galloway, 2006; Timár and Gábris, 2008). For example Sylvia and Galloway (2006) reconstructed Late Pleistocene discharge mostly based on horizontal channel parameters of the paleo-channels, whilst Lauriol et al. (2002) beside morphometry also analysed fluvial deposits to study paleo-climatic conditions. The most frequently studied channel parameters in paleo-discharge calculations are radius of curvature, wavelength, channel width and depth. Williams (1984) emphasized the regional validity of these equations, thus the equations can be used only in the same geographical environment and for the same river dimensions the equations were derived from.

In the Carpathian Basin Gábris (1986, 1995) and Timár and Gábris (2008) made paleo-discharge calculations using meander wavelength to study Holocene climate change and the discharge of scour-channels.

On the Hungarian part of the Maros alluvial fan dense paleo-channel systems can be identified, but their flow directions and pattern have not been studied in detail before, though they can provide useful information on Late Pleistocene and Holocene climate changes. The aim of the present study was to identify the abandoned channels and to determine their horizontal morphometric parameters on the Hungarian part of the alluvial fan. Our further goal was to calculate paleo-discharges using existing equations. Meanwhile, slope conditions, being potential causes of longitudinal pattern change on the alluvial fan were also investigated.

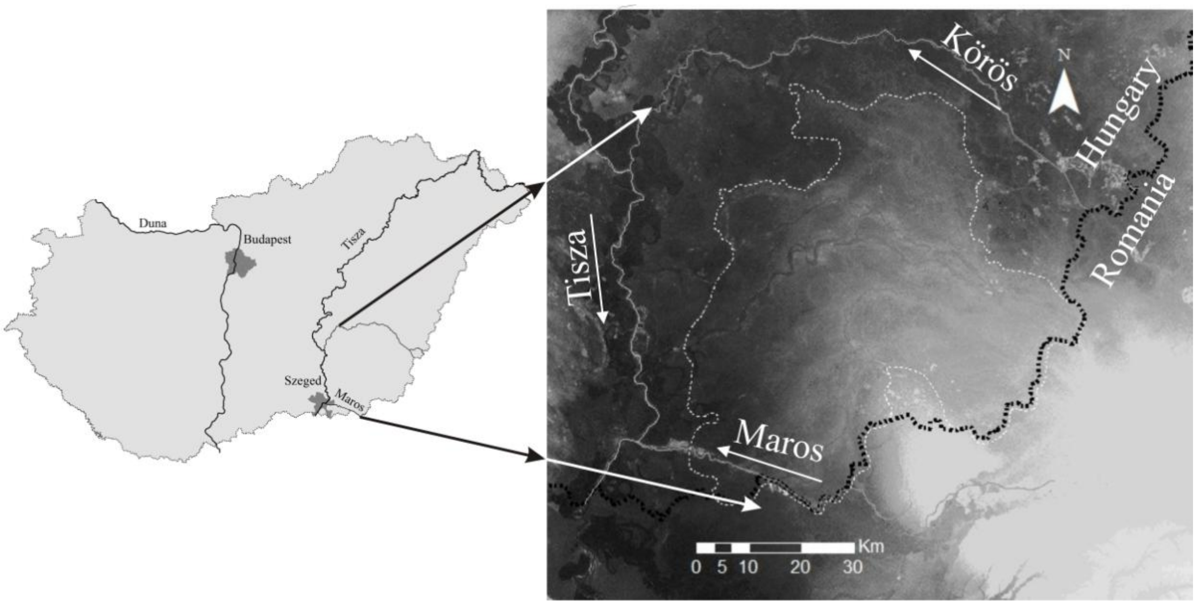

Fig. 1 The study area is on the Hungarian part of the Maros alluvial fan (SRTM model)

STUDY AREA

The alluvial fan of the Maros River has a semi-circular form with a 80-100 km radius (Mike, 1991). It is bordered by the Körös River and its floodplain in north and by the Tisza River and its incised floodplain in west. The alluvial fan stretches also to Romania and Serbia, but in the present study only the Hungarian part (3640 km^2) was studied (*Fig. 1*).

The area of the alluvial fan is affected by tectonic movements (Nádor et al., 2005, 2007). Certain parts are characterised by intensive subsidence, e.g. the Békés Basin or the Makó–Hódmezővásárhely Graben, while others are sinking more slowly, thus exhibit a relative uplifting, e.g. the Battonya–Pusztaföldvár Horst (Joó et al. 2000; Dövényi 2010). Throughout the Quaternary the direction of channels was controlled by the different tectonic activity of the above mentioned areas. The channels slid down from the rising areas and changed their directions towards the sinking basins frequently (Nádor et al., 2005), thus the body of the fan is characterised by intercalated sediment layers (Molnár, 2007). However, the history of the alluvial fan has not been cleared yet. According to Mike (1991) the evolution of the alluvial fan started in the Late Pliocene when the South Transylvanian Basin was drained by the Maros (Somogyi, 1961). However, based on the sedimentary sequences of the alluvial fan Molnár (2007) concluded that the development of the alluvial fan started just in the Early Pleistocene. According to Andó (1976) and Borsy (1989) the formation of the youngest parts started in the Late Pleistocene or in the Early Holocene. It was supported by the measurements of Gábris (1986), who revealed that the Száraz Brook was an active Maros channel or tributary channel in the Boreal Phase. According to Nádor et al. (2007) the Maros River turned to its present direction during the Middle or Late Holocene, as it is reflected by a gravel layer in the sedimentary structure of the alluvial fan. The reconstruction of the development of the alluvial fan is especially questionable along its boundaries, where the sediments of the Maros are intercalated with that of the neighbouring rivers (Molnár, 2007). The flow directions of the paleo-channels are not well known either. According to Márton (1914) the Maros had no well-defined channel until the Pleistocene, as it was split into several secondary channels. The frequent changes of river course were also emphasized by Andó (2002), who mentioned four main flow directions during the Pleistocene. According to Mike (1991), the Holocene Maros channel shifted on the alluvial fan frequently, but in each case at first it turned to north-east then towards to the south. However, Somogyi (1961) supposed a similar drainage network like nowadays at the end of the Pleistocene. The paleo-discharge was determined just for a limited number of paleo-channels (Gábris, 1986) and it showed that in the Boreal Phase mean discharge was similar (160 m^3/s) to the present-day discharge of the Maros (Fiala et al., 2006).

The surface of the alluvial fan is covered by loess, sandy loess sand fluvial sediments containing silt and clay (Dövényi, 2010). The alluvial fan is rich in fluvial forms, dominated by paleo-channels, scour-channels and cut-offs, though in some places sand deposits were blown out and dunes were formed. The Battonya Horst is poor in fluvial forms, as it lies more than 3 meters above the alluvial fan surface and is mostly covered by loess.

METHODS

To identify the fluvial forms on the Maros alluvial fan a digital elevation model (DEM) was created using 1:10,000 scale topographical maps under ArcGIS 10. software with a horizontal resolution of 2 m. Based on the DEM, the paleo-channels, mid-channel bars and point-bars were identified and used for determining channel patterns (braided, meandering, anastomosing and misfit) following the classification of Leopold and Wolman (1957) and the definitions of Rosgen (1996). Considering the pattern and the spatial distribution of forms on the surface the alluvial fan was divided into 13 paleo-channel zones. In case of the braided paleo-channels their width (W), channel length (L_c) and valley length (L_v) were measured, and their sinuosity (S) (channel length / valley length) was calculated. In case of the meandering and misfit paleo-channels the radius of curvature, (R_c), half-wavelength (L) and chord-length (H) were determined. The radius of curvature was deriveted from the largest circle fitting at least at three points to the centre-line of the meander. Half-wavelength was measured between two inflection points along the centre-line (the inflection point was understood as the mid point of the straight section between two bends). Chord-length was equal to the straight distance between two inflection points. Laczay (1982) classified river bends based on the ratio of half-wavelength and chord-length (β = L/H), and defined various development phases. Gábris (1986) suggested that from Laczay’s classes only the so called developed and well-developed meanders should be used in paleo-discharge calculations. Finally, palaeo-discharges were calculated applying the regional equations developed by Sümeghy and Kiss (2011).

During the research we were regularly faced to the problem of sudden pattern changes along the paleo-channel zones. To resolve this phenomenon, the slope conditions of the alluvial fan were also studied. The alluvial fan was divided into three belts (axial, central and distal) according to the basic morphological units of such forms (see Rachocki, 1981). The slope of belts was determined along radial lines following the paleo-channel zones. To study slope variations in detail, the slopes of the floodplain and the channel were also measured. Floodplain slope was determined along the paleo-channels, ca. 10 m away from banklines, whilst channel slope was defined along the centre-line of the paleo-channel. The entrenchment ratio was calculated by dividing the floodplain slope with the channel slope.

RESULTS AND DISCUSSION

Morphology of the paleo-channels

Paleo-channels appear in almost continuous zones (13) with well-defined boundaries (*Fig. 2)*. The mean length of paleo-channel zones is 42 km and the adjacent floodplain sections are 4.1-8.7 km wide in average. Zone No. IX has the largest territory (474 km^2), while zones No. II and VI are the smallest (ca. 99 km^2). Each paleo-channel within a zone has a typical channel pattern, though in some cases pattern-changes could also be detected (*Table 1-2)*. Two of the 13 zones (No. III and IX) are dominated by braided channels, though occasionally meandering sections do also appear. Five zones have only meandering paleo-channels, two of them are misfit (No. I and V). Three zones (No. XI, XII and XIII) are dominated by meandering channels with short braided sections. The remaining three zones (No. II, VII and X) exhibit an anastomosing pattern (although sub-channels are meandering, due to uncertain bifurcation points these were not analysed further on).

Table1 Morphometric parameters of braided paleo-channels
(A: area, W: channel width, L_c: channel length, L_v: valley length, S: sinuosity)

Zone		A (km^2)	Channel pattern	W (km)		L_c (km)	L_v (km)	S
				Min	Max			
III.	a	164	dominantly braided	2.3	4.4	62.2	50.2	1.24
	b	146		1.8	3.5			
IX.		474	dominantly braided	1.1	3.4	51.1	43.5	1.17
XI.		452	dominantly meandering	0.8	3.0	13.0	11.8	1.10
XII.		213	dominantly meandering	0.8	5.1	20.4	18.0	1.10
XIII.		375	dominantly meandering	1.6	2.5	12.0	11.6	1.04

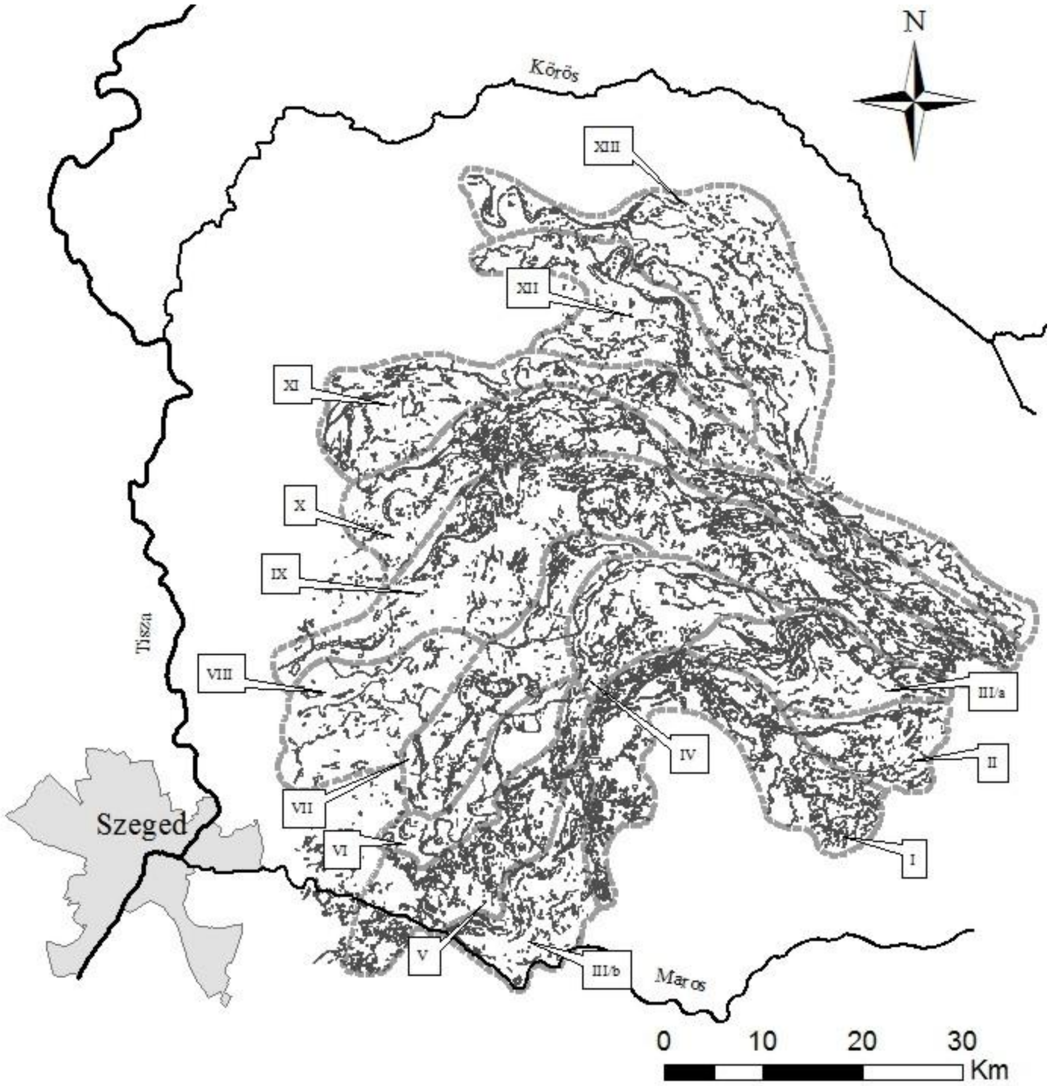

Fig. 2 Paleo-channel network on the Maros alluvial fan. The paleo-channel zones are indicated by Roman numbers (I-XIII). In the lack of absolute age data, at the present stage of the research numbering is based on the morphologically presumed succession of channels

Table 2 Mean values of horizontal channel parameters in the meandering paleo-channel zones and at subordinate meandering sections (R_c: radius of curvature, L: half-wavelength, H: chord-length)

Zone	Area (km^2)	Zone pattern	Number of analysed meanders	Horizontal parameters (m)		
				R_c	L	H
I.	296	meandering	42	202	845	519
		misfit	81	106	340	229
IV.	164	meandering	15	312	1201	710
V.	207	meandering	30	208	1008	435
		misfit	137	67	228	147
VI.	99	meandering	14	528	2175	1246
VIII.	187	meandering	13	478	1472	1113
IX.	452	dominantly braided	14	587	2394	1402
XI.	348	dominantly meandering	55	349	1331	807
XII.	213	dominantly meandering	18	297	958	712
XIII.	375	dominantly meandering	32	656	1775	1390

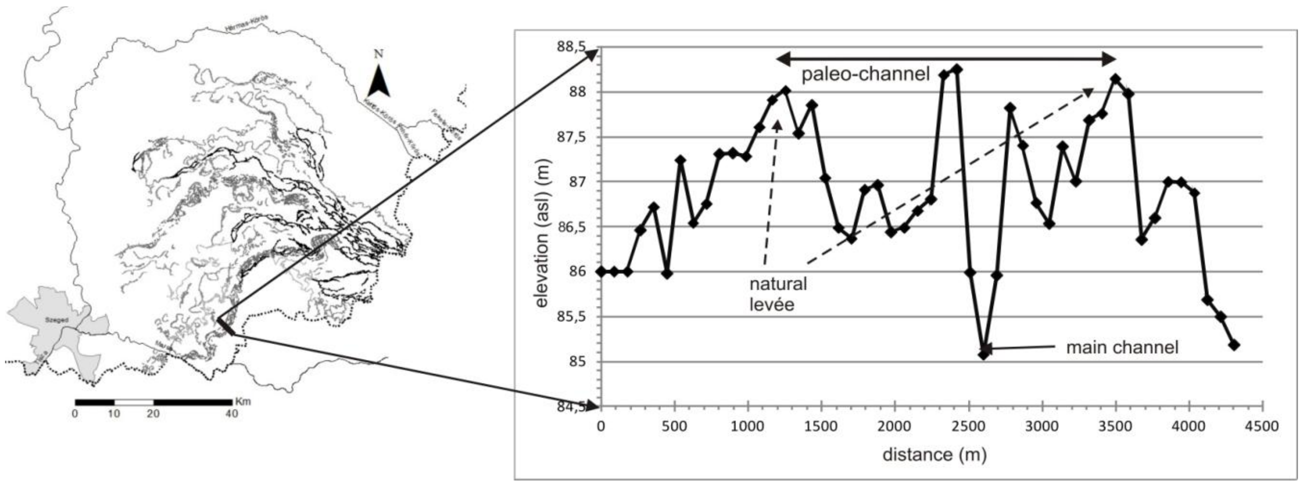

Fig. 3 Cross-section of a braided channel in zone No. III/b

The minimal width of ***braided channels*** varies between 0.8 and 2.3 km, while their maximal width is between 2.5 and 5.1 km *(Table 1)*. The two dominantly braided channel zones (No. III and IX) stretch from the apex of the alluvial fan to its distal, western border, therefore their channel length is between 51.1 and 62.2 km, whilst the length of their valley is 43.5 and 50.2 km respectively. The sinuosity of the braided channels is between 1.0 and 1.2, which fits well to the Leopold and Wolman's classification (1957), having a 1.5 value as a threshold. The forms of the two dominantly braided channels are almost entirely recognisable, though some sections have been reworked by younger meandering and misfit channels. Their natural levees are approximately 1.0-1.5 m higher than the surrounding floodplain surface *(Fig. 3)*.

On the northern part of the alluvial fan, which is dominated by meandering zones, three braided paleo-channel remnants were found. They are only 12.0-20.4 km long and their valley length is 11.6-18.0 km. Certain sections were reworked by meandering channels, thus the banks and the bars partially disappeared.

Meandering channels occupy the southern and western parts of the alluvial fan. Concerning the meandering zones the value of mean R_c varies between 202 and 656 m, L is 845-2394 m, while H is between 435 and 1402 m (*Table 2*). However, the number of the analysed bends was quite different in each zone. The fewest developed and well-developed meanders (in all 13) were identified in paleo-channel zone No. VIII, while there were 55 such meanders in the zone No. XI. The greatest horizontal channel parameters were measured in zone No. XIII (R_c=656 m), and in zone No. IX (L=2394 m, H=1402 m). The smallest mean ratio of curvature (R_c=202 m) and half-wavelength (L=845 m) was measured in zone No. I, while the smallest chord-length (H=435 m) was in zone No. V *(Fig. 4)*. Thus, in case of each parameters the difference in mean values is almost threefold (2.8-3.2).

Decreasing discharge resulted the development of ***misfit channels*** (zone No. I and V), which then reworked the material of the original channel. The mean half-wavelength of misfit paleo-channels is 228-340 m, while the mean chord-length is 147-229 m. The mean value of the radius of curvature is varying between 67-106 m. Values related to the original meanders are 2-4.5 times greater than that of the misfit bends *(Table 2)*.

As according to Gábris (1986) developed (β=1.1-1.4) and well-developed (β=1.4-3.5) meanders are the most appropriate for paleo-discharge calculations, the development phase of the bends was also calculated. Concerning mean values two out of the nine meandering paleo-channel zones (No. XII and XIII) fall into the developed class, while the remaining seven zones are well-developed *(Table 3)*. For all zones the paleo-discharge calculations were carried out using the mean horizontal channel parameters and the formerly created regional equations (Sümeghy and Kiss, 2011).

The greatest mean bankfull discharge was calculated for paleo-channel zone No. IX (2655 m^3/s). This value equals to the present day flood discharge (2400-3200 m^3/s) of the Tisza River (Sándor, 2011) and slightly higher than the present day peak discharge (2420 m^3/s) of the Maros River (Sipos, 2006). On the alluvial fan there are two other paleo-channels with bankfull discharges above 2000 m^3/s (zone No. VI and XIII). Four paleo-channels (No. IV, VIII, XI and XII) have a mean bankfull discharge around 800-2000 m^3/s, which is similar to the mean discharge of the Tisza (800 m^3/s, Tímár, 2003). In zone No. I. original meanders could have four times greater bankfull discharge (493 m^3/s) than younger, misfit channels (119 m^3/s). The difference is even greater in case of zone No.

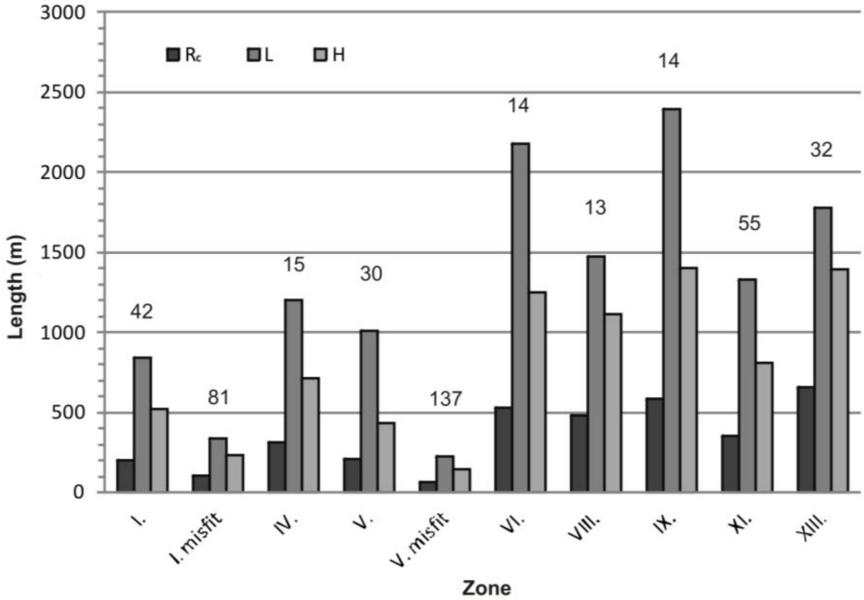

Fig. 4 Mean radius of curvature (R_c,), half-wavelength (L) and chord-lendth (H) values of the meandering paleo-channel zones and the number of the analysed meanders

V., where the original discharge was ca. 508 m^3/s, while the discharge of the misfit channel was only 27 m^3/s. In these cases the earlier discharge of the paleo-channel zones is similar to the present-day bankfull discharge of the Maros (680 m^3/s, Sipos, 2006) and lower than the mean discharge (600 m^3/s) of the Tisza (Sándor, 2011). The bankfull discharge values of the misfit channels are close to the minimum discharge (31 m^3/s) of the present day Maros (Sipos, 2006).

Slope conditions on the alluvial fan

The height difference between the highest (103 m asl) and lowest (77 m asl) points of the alluvial fan is almost 30 m. The greatest slope was found in the central part of the fan (mean slope: 31.0 cm km^{-1}), whilst slightly lower values were experienced in relation with its axial (23.8 cm km^{-1}) and distal parts (24.9 cm km^{-1}, *Fig. 5)*. These parameters reflect the normal radial profile of an alluvial fan (Rachocki, 1981) referring to coarse sediment deposition in the central zone of the alluvial fan (Lecce, 1990).

Our aim was to find relationship between slope and channel pattern. The mean channel slope of meandering zones is 12.3 cm km^{-1} and of the braided channels is slightly higher, 17.6 cm km^{-1}. Therefore, in the central part of the alluvial fan, which has the greatest surface slope, more braided channels or braided channel sections were formed than in the distal zone, which is characterised by mostly meandering channels.

Table 3 The β-values (L/H) refereeing to the development phase of bends and the calculated mean bankfull discharges (Q_{bf}). Discharge calculations were based on the radius of curvature (R_c), half-wavelength (L), chord-length (H)

Zone	β	Bankfull discharge (m^3/s)			
		R_c	L	H	Q_{bf} mean
I.	1.68	649	423	405	493
	1.46	225	70	61	119
IV.	1.80	1154	765	770	896
V.	2.35	673	570	281	508
	1.55	58	13	10	27
VI.	1.73	2199	2087	2376	2220
VIII.	1.40	1919	1076	1899	1631
IX.	1.80	2497	2462	3006	2655
XI.	1.58	1327	908	998	1078
XII.	1.30	1084	524	774	794
XIII.	1.29	2853	1474	2955	2427

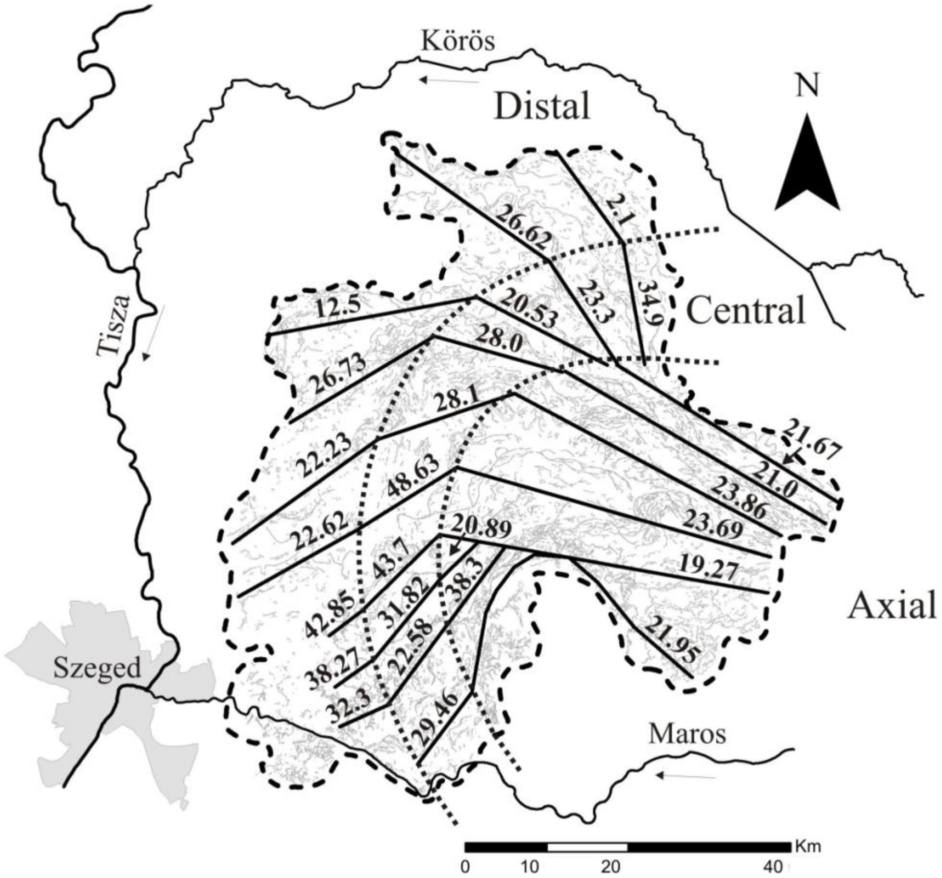

Fig. 5 Slope calculations were made along radial lines following the paleo-channel zones in the axial, central and distal parts of the alluvial fan (cm km^{-1})

The entrenchment ratio of braided paleo-channels is between 1.0 and 1.4, while meandering paleo-channels are more incised (entrenchment ratio: 1.5-2.0), and in case of misfit paleo-channels the ratio is above 2.0. Comparing the longitudinal slope profiles it becomes clear that in case of the braided paleo-channels longitudinal profiles of the floodplain and of the channel are almost parallel. In contrarily, the profiles of the meandering and misfit channels are convergent towards downstream. This can be explained by the tectonic uplift of the apex part of the alluvial fan or by an intensive accretion on the distal parts.

The relationship between channel slope and the channel pattern changes is the most conspicuous in zone No IX. The channel pattern is braided in the axial part of the alluvial fan, meandering in the central part and in the distal part is braided again. These changes are reflected in channel slope as well. The channel slope is 22.4 cm km^{-1} in the eastern part, 19.2 cm km^{-1} in the meandering zone and 25.6 cm km^{-1} where the pattern changes and braided pattern appears again. The pattern change shows similar characteristics in other zones as well, thus the slope of braided channels is usually higher than that of the meandering ones.

CONCLUSIONS

The surface of the Hungarian part of the Maros River alluvial fan was divided into 13 paleo-channel zones based on the channel pattern and morphometric parameters of the identified paleochannels. These zones showed braided, meandering, misfit and anastomosing patterns, though in some zones have a mixed character.

Braided paleo-channels are exceptionally large, their mean width varies between 1.4 and 3.7 km. They are bordered by 1-1.5 m high natural levees, and their channel is dissected by bars. The sinuosity of the braided channels is between 1.0 and 1.2, being under the threshold (1.5) of Leopold and Wolman (1957). The longitudinal profile of the floodplain and the braided paleo-channels are almost parallel (entrenchment ratio: 1.0 and 1.4). These channels mostly appear in the central zone of the alluvial fan, where the surface slope is the highest (its mean value is 31.0 cm km^{-1} and the maximum slope is 49.0 cm km^{-1}), thus these channels could play important role in the sediment transportation and aggradation of the alluvial fan. Similar braided pattern can not be found nowadays in the Carpathian Basin, but they were more abundant in former geological times. Some braided paleo-channels were found on the Romanian part of the Maros's alluvial fan with very similar channel parameters. Braided paleo-channel generations were also identified in the Middle Tisza Region (Gábris et al., 2001) dating back to the Late Glacial Maximum and to the Younger Dryas periods. However, on the alluvial fan of the Hernád and Sajó Rivers braiding channel networks were not identified on the surface despite of the frequent channel changes (Nagy and Félegyházi, 2001; Nagy, 2002).

Meandering channels on the alluvial fan have a highly variable size. Values horizontal morphometric parameters were the lowest in the case of misfit channels (Rc=67-106 m, L=228-340 m, H=147-229 m), thus they had the smallest bankfull discharge (27 and 119 m^3/s). Nevertheless, much larger meanders do also exist on the alluvial fan (Rc=202-656 m, H=845-2394 and L=435-1402 m) with bankfull discharges between 490-2650 m^3/s. The meandering paleo-channels are mostly located in the central and distal parts of the alluvial fan. Their channel slope varies around 12 cm km^{-1}. The entrenchment ratio of meandering channels is between 1.5 and 2.0, referring to slight incision. Floodplain and channel slopes are convergent downstream referring to tectonic uplift at the axial part of the alluvial fan or intensive downstream accumulation.

According to the calculations of Gábris (1986), the mean discharge of the Száraz Brook (in paleo-channel zone I) varied on the alluvial fan between 34 m^3/s and 307 m^3/s. In the present study the paleo-channel of the Száraz Brook was classified as a misfit channel. In zone No. I its mean bankfull discharge was 119 m^3/s, while the older meandering paleo-channel, in which it was developed, had a ca. 500 m^3/s mean bankfull discharge.

Comparing the calculated bankfull discharge to the present values of the Tisza and Maros we found that the largest meandering paleo-channels had bankfull discharges (2427-2655 m^3/s) similar to the present-day peak discharge of the Maros (1600-2500 m^3/s, Fiala et al., 2006) and the flood discharge of the Tisza (2400-3200 m^3/s, Sándor, 2011). The discharges of medium sized meandering paleo-channels (493-794 m^3/s) are around the present day bankfull discharge of the Maros (680 m^3/s, Sipos, 2004) and the mean discharge of the Tisza (800 m^3/s, Tímár, 2003). However, the smaller, misfit paleo-channels transported only 27-119 m^3/s water, comparable to the minimum discharge of the Maros (31 m^3/s, Sipos, 2006).

Acknowledgements

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AEOLIAN SURFACE TRANSFORMATIONS ON THE ALLUVIAL FAN OF THE NYÍRSÉG

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Abstract

The evolution of the Nyírség and its landforms has been widely addressed by Hungarian geographers and geologists in the past and at present as well. Early works were mainly concentrating on understanding the complex, fluvial and aeolian genetics of the territory, later more specific forms and problems were studied and revealed. By the increasing number of chronological evidence derived from stratigraphic, pollen and archaeological data and absolute dating techniques (radiocarbon and optically stimulated luminescence) the geomorphological development can be reconstructed in more detail, especially in the context of climate variations and human intervention. Our study aims to summarize and outline the times of aeolian activity, with special respect to Holocene events, on the basis of the researches carried out so far. By nowadays it is obvious that sand was moving on several occasions in the Nyírség during the Holocene subsequent to the main aeolian land formation periods of the Upper Pleniglacial and the late glacial. In the first half of the Holocene sand movement can be related to dry periods, thus aeolian activity was driven mostly by climatic factors. In the second half of the Holocene the area of land affected by wind erosion decreased and in most cases events can be associated to the activity of man. Nevertheless, climatic and anthropogenic factors could be superimposed, leading to significant local sand mobilisation.

Keywords: wind-blown sand, Holocene, climate change, human impact

INTRODUCTION

Earth scientists have studied the Nyírség for over 100 years but geological and geomorphological research can still contribute to the understanding of the surface development of the area. According to Nagy (1908), the sand of the Nyírség was placed to its current location from the alluvial fans of the Ondava, Tapoly, Laborc, Ung and Latorca rivers in the foreground of the volcanic mountains, i.e. it was transported by wind from distant areas.

Cholnoky (1910) in his work entitled “Surface of the Great Hungarian Plain” discussed the sand area of the Nyírség in detail and explained also the development of the landforms. He thought also that the sand was originated from the “debris” of the Ondava-Tapoly-Ung and Latorca and according to him the “strong northern foehn winds formed semi-bond wind-blown sand forms on the loess plateau of the Nyírség”. According to these, the wind-blown sand was formed after the formation of the loess. In his opinion the sand of the Nyírség came from a long way and this explained the roundedness of the grains. One characteristic of the semi-bond wind-blown sand is that the forms develop where vegetation cannot protect the surface. Therefore it is hard to believe that grains could have been transported for such a long distance along a surface covered partially by vegetation. Nevertheless, the idea of the renowned geographer remained for decades influencing other researches as well.

The theory of Cholnoky became outdated when Sümeghy (1944) – based on borehole data – described the Nyírség as an alluvial fan built by rivers flowing from the Carpathians and Transylvania. He also noted that no loess can be found underneath the wind-blown sand but fluvial sand supplying the material blown by winds to form wind-blown sand. His theory was proved by stratigraphic and morphological research as well.

Aeolian landforms of the Nyírség

Significant differences can be found in the descriptions of the landforms in the Nyírség. Nagy (1908) considered ripple marks, hummocks and yardangs, as he called dunes dominant. He also wrote that sand moved towards the south in the form of barchans. He noticed that there are significant differences among ripple mark surfaces and that different forms can be found south and north of the watershed in the Nyírség. Formation of the landforms, however, was explained improperly by him.

According to Cholnoky (1910), sand dunes are the forms of semi-bond sand characterised by crests elongated according to the orientation of the prevailing wind and by ripple marks in between them. In his opinion westerly winds pushed the middle section of the crests slightly to the east, however, the main direction remained.

Balla (1954) writes about parabolic blowout dunes, longitudinal dunes, Kádár type Lybian dunes, ripple marks and coastal dunes. He explains the formation of the coastal dunes by eastern and northeastern winds. In his opinion, sand blown from the beds with low water level was accumulated in the form of dunes along the shores in the dry periods of the Würm.

Kádár (1956) discussed the Nyírség as well in his work studying wind-blown sand. In this publication abandoned his earlier theory related to Libyan dunes and he termed the parabola dunes elongated along their eastern wing beside inter-dune water as a marginal dune.

Borsy (1961) carried out detailed geomorphological mapping based on field-research of several years in the Nyírség resulting in the first geomorphological map of a Hungarian sand area. He studied the size, orientation and internal stratification of the landforms drawing conclusions on the palaeo-wind conditions as well. He systemized and classified the deflation and accumulation forms and described their development in detail. He discussed the areas of the Nyírség and their aeolian landforms separately.

AGE OF THE WIND-BLOWN SAND LAYERS

The first statements related to the age of the wind-blown sand landforms in the Nyírség are written in the book of Borsy (1961). When studying the surface of the Nyírség he assumed Holocene sand movements based on the more intact landforms greater in size. He also noticed the fossil soils dividing the dunes, and identified the loess layers as well. He reckoned the formation of the wind-blown sand above the loess layer took place in the Hazel phase when the climate was drier and warmer than today. He changed his views regarding sand movement in the Holocene when ^{14}C age determinations enabled the more accurate calculation of wind-blown sand movement periods (*Fig. 1*).

Radiocarbon (^{14}C) age data

The first results were obtained when the strata of the sand quarry in the western outskirts of Aranyosapáti were studied where two wind-blown sand strata of different age are divided by a loess layer (*Fig. 2*). Charcoal remnants in the fossil soil developed on the loess made ^{14}C age determination and thus the more accurate definition of the wind-blown sand movement periods possible.

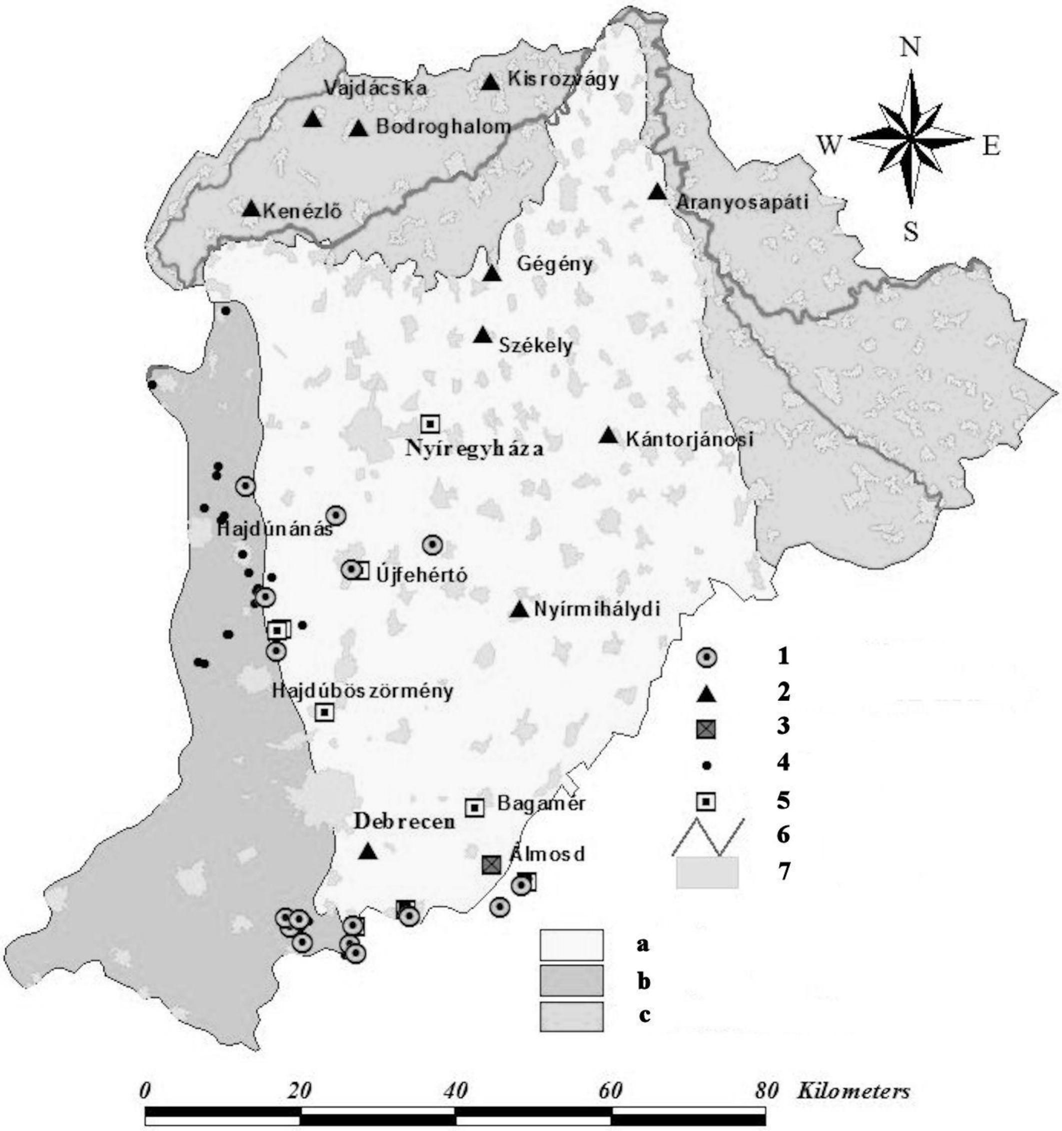

Fig. 1 Location of boreholes and outcrops in the Nyírség and in the marginal areas of the alluvial fan
1:pollen boreholes; 2:radiocarbon; 3:outcrops; 4:boreholes; 5:OSL; 6:rivers; 7:settlements; a:Nyírség; b:Hajdúhát; c:Upper-Tisza-region

Age of the fossil soil is 12900±500 years BP, i.e. it was formed in the first milder climate of the late glacial. The wind-blown sand below the loess layer was formed before this date from the fluvial sand of the alluvial fan. Based on the radiocarbon data obtained from the research carried out in the Nyírség, Bodrogköz and the Danube-Tisza Interfluve, the first major accumulation of wind-blown sand occurred in the Upper Pleniglacial and it was followed by several in the Dryas (Borsy et al., 1981, 1985; Lóki et al., 1993).

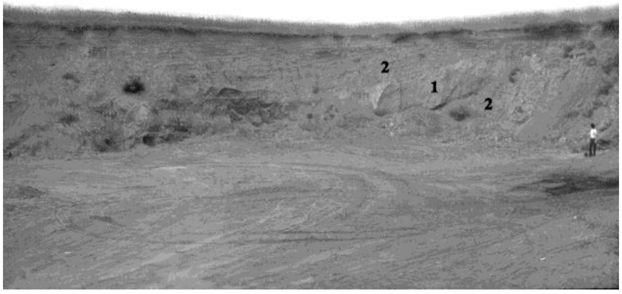

Fig. 2 Sand quarry in the outskirts of Aranyosapáti (1: loess, 2: wind-blown sand)

Based on the data gained by the research in the 1980s it can be stated that the landforms of Hungarian wind-blown sand areas were formed in the dry periods of the late glacial and sand moved only in a limited area primarily due to anthropogenic effects in the Holocene.

A humus layer at small depth underneath the surface (*Fig. 3*) is traceable in numerous outcrops in the Nyírség that is the continuation of the surface soil frequently. Several scientists (Marosi, 1967; Borsy, 1961; Borsy and Lóki 1982; Lóki, 2003) explained the burial of these soils by wind erosion as a result of deforestation in the 18th century. This correct statement has to be completed by that – on the basis of historical data – deforestations took place in the early 19th century as well in order to increase the ratio of arable lands, thus wind erosion could have made its influence felt at that time as well.

In the course of recent research (Lóki et al., 2008) new outcrops were found in which fossil soil divides the wind-blown sand strata. These, however, are thinner (*Fig. 4*) than the soils formed in the late glacial (*Fig. 5*). This suggests shorter period suitable for soil formation. In the subsequent dry period soil moved again. Radiocarbon age data of soils prove sand movements in the Holocene. Age of the fossil soil in the outcrop near Gégény is 3740±50 years (BP) and it is covered by 4 m of wind-blown sand. Age of the fossil soil in the sand quarry at Kántorjánosi is 9300±50 years (BP) and this is also covered by 4 m of wind-blown sand (*Fig. 6*).

Fig. 3 Young fossil soil in a wind-blown sand outcrop in the Nyírség

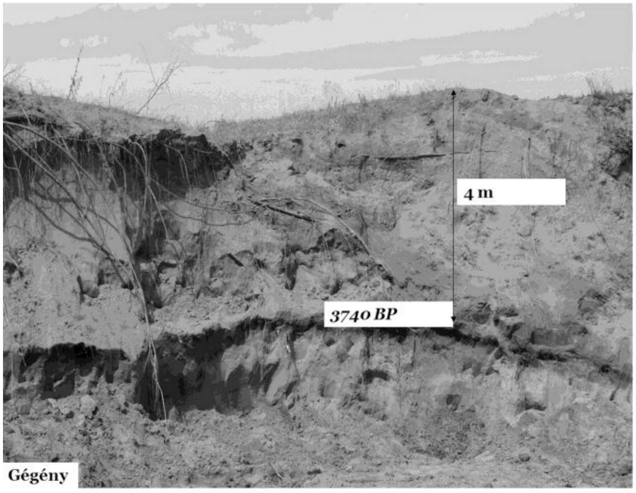

Fig. 4 Position and age of the fossil soil horizon (sand quarry at Gégény)

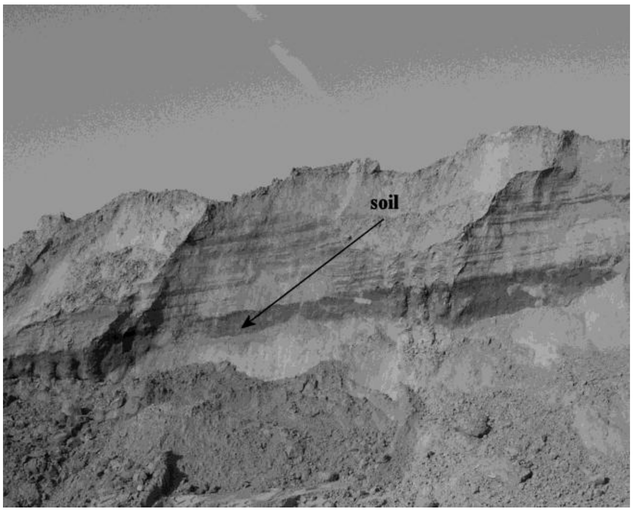

Fig.5 Late glacial soil east of Nyíregyháza

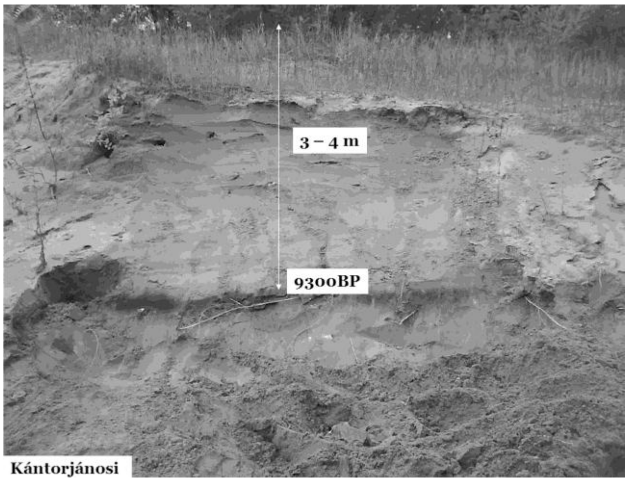

Fig. 6 Position and age of the fossil soil horizon (sand quarry at Kántorjánosi)

The two soil horizons of different age suggest that sand moved at several times in the Holocene and that wet and dry periods alternated. However, outcrops were also found in the Nyírség in which no fossil soil was detected in the 4-6 m thick wind-blown sand on top of the fluvial sand (*Fig. 7*). Age of these strata was determined by OSL measurements

Fig.7 Wind-blown sand outcrop with strata showing different grain-size distribution

OSL age data in the Nyírség

In recent years Kiss and co-workers carried out OSL measurements on samples from the Nyírség. These measurements are important because they enable the determination of the age of sand strata without soils as well. Their research at study areas near Bagamér and Erdőspuszta revealed wind erosion during the Atlantic and Subboreal phases (Kiss and Sipos, 2006; Kiss et al., 2008). Sand movement in the Atlantic phase in the Nyírség was already presumed by Borsy (1980), however, at that time measurements were not available to prove that.

Data were obtained from research in other Hungarian wind-blown sand areas in the last decade (Ujházi, 2002; Ujházi et al., 2003). Ujházi (2002) showed Upper Atlantic sand movement above the older Dryas applying luminescence measurements in outcrops at Dunavarsány. Research carried out in the Danube-Tisza Interfluve (Lóki and Schweitzer, 2001; Nyári et al., 2006; Kiss et al., 2008; Sipos et al., 2006) verified sand movements at the beginning of the Subatlantic phase, in the Iron Age and at the time of the great migrations as well. All these data indicate that aeolian land formation prevailed at several times in the Holocene.

Stratigraphic and pollen data

Changes in the stratification of accumulation forms, archaeological findings in the outcrops and alterations of the pollen content of the strata yield more data for making the reconstruction of aeolian surface development more accurate.

Considering the stratification of the wind-blown sand outcrop near Gégény mentioned above it can be stated that the accumulation of the 4 m thick wind-blown sand covering the Subboreal fossil soil was followed by another sand movement. Succession in the eastern wall of the dune (*Fig. 8*) supports this as sloping strata are covered by almost horizontal wind-blown sand strata. These are made variable by iron pan layers. Strata series with different orientations suggest changes in wind directions. The two differing strata series are not divided by a younger soil horizon because the conditions of soil formation were missing on the loose and dry sand surface.

Fig. 8 Strata series with different orientation in the sand quarry at Gégény

The stratigraphic analysis included the surfaces covered by shroud of sand in the marginal areas of the Nyírség and the river beds buried partly by wind-blown sand as well. Archaeological findings helped determining the age of sand movement (*Fig. 9*) in the strata of outcrops in the areas covered by shroud of sand (Félegyházi and Lóki, 2006).

Based on our research so far it can be stated that the sand sheet covers in the southern marginal areas of the Nyírség were formed in several periods. Sedimentological and palynological analyses verified clearly the accumulation of strata related to shroud of sand at the end of the Pleistocene and in the Holocene. Analysis of the Upper Pleniglacial Selaginella containing bog strata exposed in the outskirts of Álmosd and Kokad and south of Hajdúbagos suggests the settlement of bog pine indicating dry summer. In the periods of drought the grass vegetation bonding the surface on the dunes became dry and the sand started to move. Further movement of the sand was impeded only by the swamps and bogs developed in the marginal areas in wetter periods therefore wind spread only little sand onto the bogs and swamps.

Fig.9 Strata series in the borehole drilled near Hosszúpályi (a:wind-blown sand, b:findings of Sarmatian Age, c: findings of Bronze Age)

Stratigraphic and palynological analyses of boreholes drilled in the abandoned riverbeds in the alluvial fan of the Nyírség contributed to the more accurate reconstruction of its land development. One borehole was drilled in the depression starting from Hajdúhadház through Újfehértó and Nyíregyháza up to the Small Tisza, while another borehole is located in the bed remnant detectable from Nyíradony to Rétköz.

In the 550m deep borehole drilled in the bed remnant in the vicinity of Újfehértó the bottommost sediment that yielded pollen was found in the stratum between 520 and 480 cm (*Fig. 10*). The 20% ratio of floating red-grass and galingale (*Cyperaceae*) among the pollens indicate lacustrine conditions with low stagnating water. The 10% of spikemoss (*Selaginella*) among the pollens indicate clearly the cold, boreal climate of the Pleistocene. Dry terraines of the area were covered by steppe meadows with pine groves. The pine forests were composed of Swiss pine (*Pinus cembra*), Scots pine (*Pinus sylvestris*), spruce (*Picea*) that are characteristic for the taiga and have always been rich in the forests of the wetter taiga. Wetter climate is also supported by the presence of willow (*Salix*). Similar pollen composition was detected in the strata of the Fehér lake at Kardoskút and in the older strata of the Nagymohos bog at Kelemér from the interstadial of the Upper Plenniglacial the of which was determined to be 26000–24500 years BP (Magyari, 2002; Magyari et al., 2000; Sümegi et al., 1999; Járainé Komlódi, 2000).

At the depth between 480 and 430 cm one pollen shows a not interpretable sandy, silty sediment layer. This accumulation did not help the preservation of pollen grains. Micro-stratigraphic analyses suggest slight wind-blown sand movements that can be explained by a short dry, cold period. Sand movement of this age has not been identified in the Nyírség so far but research carried out in the Danube-Tisza Interfluve (Krolopp et al., 1995) suggests sand movement of similar age (25200±300 BP). Magyari (2002) indicates steppe–tundra vegetation between 24500 and 22700 years BP when characterising the vegetational phases of the northeastern part of the Great Hungarian Plain.

Pollen content of the sediments between 430 and 360 cm suggests wetter climate again. Pollen composition can be characterised by a mixed foliage taiga forest. This milder climate can be shown at several places in the Carpathian Basin (Járainé Komlódi, 1969, 1970; Borsy and Félegyházi, 1991; Sümegi and Magyari, 1999; Félegyházi and Tóth, 2003, 2004) characteristic for the period between 22700 and 21400 years BP. Pollens upwards in the sediment series indicate that the climate became more-and-more continental gradually. This fits well in the vegetational phase prevailed between 21400 and 18500 years BP (Magyari, 2002). Severe winters and dry, warm summers destroyed woodland vegetation and only grasses survived among herbaceous plants. This is indicated in the pollen spectra by that the ratio of pine pollens fall below 10% while that of grass species (*Gramineae*) increases up to 60%.

After this pollen accumulation was terminated, the bed was covered by a 200 cm thick wind-blown sand stratum. Vegetation composed of cold steppe grasslands, then it was dried out as well and the surface of the alluvial fan was shaped by cold, sometimes stormy winds. Formation of this sediment stratum took place around18500–14600 years BP.

The sediments containing pollen appear again at the depth of 150 cm. Pollen distribution is dominated by pine in 80% regarding arborescent vegetation. The remaining 20% is presented by deciduous trees and herbaceous vegetation with diverse taxon composition. At the time of the deposition of this stratum lacustrine conditions characterised this part of the bed.

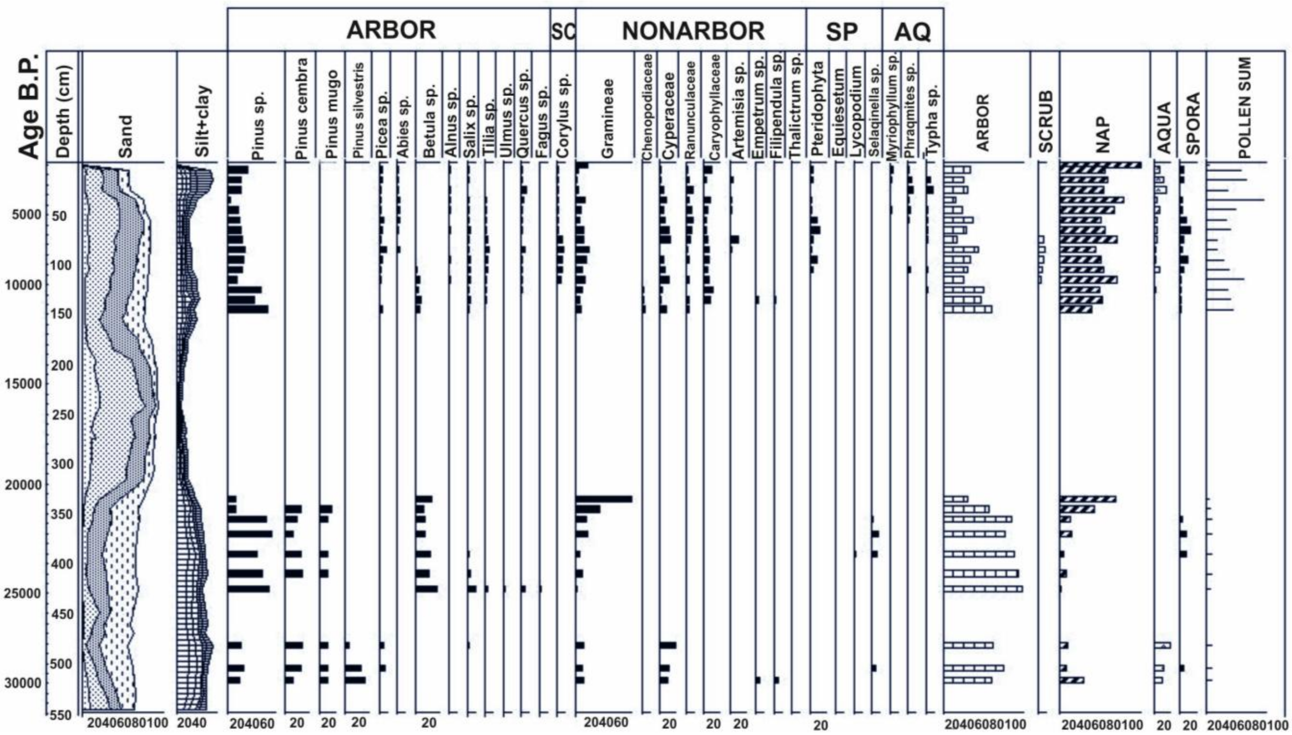

Fig. 10 Palynological and grain-size distribution diagrams of the sediments from the bed remnant in the vicinity of Újfehértó

With the gradual warming of the climate deciduous trees – birch (*Betula*) first, then alder (*Alnus*), willow (*Salix*), linden-tree (*Tilia*) and oak (*Quercus*) appeared, however, Scots pine (*Pinus sylvestris*) was still present on the sand. Hazel (*Corylus*) was present in the underwood of oak forests as well. This sediment stratum represents the forest advancement phases of the post-glacial time and the start of the early Holocene.

Results of the palynological analyses of the cores from the borehole drilled in the other bed remnant between Érpatak and Nagykálló are the same as those of the borehole in the vicinity of Újfehértó. Thickness of the late glacial wind-blown sand is almost 200 cm at both locations.

It can be also concluded from the analysis results of the samples from the two boreholes that at the time of formation of the lower sediment series lacustrine conditions prevailed instead of the previous fluvial conditions, i.e. the rivers accumulating the alluvial fan did not flow through the alluvial fan due to the depressions forming in the foreground of the mountains. In the dry period winds buried the abandoned beds completely at places or partially elsewhere. Only smaller sections of these beds can be observed nowadays.

CONCLUSION

In the development of the surface of the alluvial fan in the Nyírség wind also had an important role apart from water. Rivers accumulated vast amount of sandy sediments in the alluvial fan accumulating in the foreground of the mountains to the Körös rivers. Flow direction of these rivers changed several times during the accumulation of the alluvial fan. Winds re-worked sandy sediments on the surface in the dry periods of the Würm then fluvial sediment series covered the aeolian deposits in the wetter periods. Although there is no direct evidence for this in the Nyírség but the presence of wind-blown sand in the deeper strata of the alluvial fan was detected with the help of electron-microscopic analysis of sandy sediments from cores drilled in the Danube-Tisza Interfluve (Borsy et al., 1987). Applying electron-microscopic analysis on the sandy sediments of the alluvial fan of the Nyírség in the future the aeolian or fluvial development of the strata could be determined. Stratigraphic and morphological research of the last the decades focused only on the near surface (< 20 m) strata of the alluvial fan. Deposition and re-working of these sediments took place in the last 30000 years.

Based on the research carried out so far it can be stated that sand was in motion in the Nyírség at several times in the Holocene as well after the aeolian land formation in the Upper Pleniglacial and the late glacial determined in the 1980s (*Table 1*). In the first half of the Holocene sand movement took place in the Boreal phase and in dry periods of the Atlantic phase thus the aeolian shaping of the land can be explained by climatic reasons. In the second half of the Holocene the land shaping effect of wind probably prevailed in a much smaller area and it was associated with the activity of man. Naturally, periods of drier climate had an important role in the aeolian land shaping due to anthropogenic effects.

Table 1 Chronological order of Holocene sand movements

Climate periods of Europe	**Archaeological chronology**	**Geochronology**		**Pollen climate**	**Sand movement (BP) climatic+anthropogenic**
		old	**new**		
2000 AD		Subatlantic	Subatlantic	0	
	1700 AD modern era			Beech II	Bagamér 230-90 OSL (Kiss, T.) Bagamér 430-350 OSL (Kiss, T.)
1300-1700 AD Little Ice Age					
					Erdőspuszta OSL 900-1000 (Kiss, T.)
600 AD drying maximum	deserta Avarorum				
	450 AD time of the great migra- tions				Bagamér 1370, 1100, 960 OSL
	1 AD Roma, Sarmatians, Dacians			BP. 2000	Hosszúpályi Sarmatian pot (Lóki, J.)
				Bükk I	
300 BC climate optimum					
		2600 BP	2500 BP		Bagamér 2050 OSL(Kiss, T.)
	850 BC Iron Age Celts, Veneti	Szubboreal	Szubboreal s		Bagamér 2450 OSL(Kiss, T.)
1200 BC drying maximum	migration of marine cul- tures, defeat of Troy, Dorians, Hebrew				Gégény 14C 3740 (Lóki, J.)
2100 BC climate optimum	Palace farming in Minos, Crete				
	2800 BC Bronze Age				Hosszúpályi Bronze Age pot remnant (Lóki, J.)
3000 BC drying maximum	Nile valley, Mesopotamia irrigation agriculture			BP. 5000	
				Oak	
		5600 BP	5700 BP		Bagamér 6000 OSL (Kiss, T.)
	4400 BC Copper Age	Atlantic	Atlantic		Bagamér 7070 OSL (Kis,s T.)
		7300 BP			
		Boreal			
	6000 BC Neolithic			BP. 8000	
			8300 BP	Hazel	
			Boreal	BP. 9000	
				Silver birch	Bagamér 9200 OSL (Kiss, T.) Kántorjánosi 14C 9300 (Lóki, J.)
		9600 BP			
		Preboreal			South Nyírség pollen (Félegyházi, E. and Lóki, J.)
		10200 BP			
		Dryas	11200 BP		
	10000 BC Mesolithic		Dryas		

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RECONSTRUCTION OF PALAEO-HYDROLOGY AND FLUVIAL ARCHITECTURE AT THE OROSHÁZA PALAEO-CHANNEL OF RIVER MAROS, HUNGARY

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Abstract

Several studies have addressed the impact of climate change and tectonic activity on fluvial systems. When investigating these systems palaeo-hydrological and geomorphological data on abandoned channels can yield valuable results. The main aim of our work was to reconstruct morphological conditions at the Orosháza palaeo-channel and to estimate the bankfull discharge which characterized the channel during its formation. There are several equations predicting bankfull discharge on the basis of planform parameters, but these only work for meandering rivers. In case of braided channels flow reconstruction can only be made by using cross-sectional parameters. The Orosháza palaeo-channel provided the means of a comparative analysis in this respect. By a sudden pattern change both meandering and braided reaches, supposedly having a very similar bankfull discharge, could be simultaneously studied. Planform parameters and present cross-sections were determined on the basis of a high resolution DEM, while original cross-section parameters were assessed using sedimentological and geophysical methods. Based on sedimentological data, channel pattern transition was mainly driven by intensive bedload accumulation at the edge of the Maros Alluvial Fan (MAF). Slope differences could not be evened out due to an avulsion close to the apex of the fan. Concerning discharge calculations a good agreement was found between a region-specific planform based equation and the cross-section based Grauckler-Manning equation. Values determined for the braided and meandering reach were also in a good correspondence. Consequently, the presented approach is suitable to determine the discharge of other braided palaeo-channels on the MAF and elsewhere.

Keywords: morphological reconstruction, hydrological reconstruction, sedimentology, geophysics, discharge equations

INTRODUCTION

The reconstruction of fluvial systems is of key importance if the climatic variations or tectonic development of an area is investigated. Therefore in the past decades numerous studies have been made to investigate past fluvial architecture, morphology and sedimentology at various locations (e.g. Dury, 1976; Sridhar, 2007; Timár and Gábris, 2008). When studying ancient fluvial systems one of the basic approaches is to allocate and investigate abandoned channels still detectable on the surface. These provide information primarily on Late Pleistocene and Holocene changes in the environment (Carlston, 1965; Bridge, 2003). Based on the morphometry and sedimentary structure of these channels various conclusions can be made on the discharge of the forming river and the quality and quantity of its sediment. These in turn might provide information to the development of contemporary topography and palaeo-climatic conditions. The interpretation of surficial and shallow deposits is either based on in situ observations, laboratory experiments or geophysical methods (Bridge, 2003; Sridhar, 2007).

The fluvial network of the Pannonian Basin has been affected in the past by climate variations and different rate of tectonic subsidence and uplift processes, the effect of base level change can be regarded negligible (Bridge, 2003; Timár et al., 2005; Gábris and Nádor 2007; Nádor et al. 2007). Tectonic processes are primarily important in determining the gradient and the direction of the flow of rivers (Bridge, 2003; Laure, 2008). As the tectonic development of the Pannonian Basin has been very complex, the reconstruction of the changes in the fluvial network since the regression of the Pannonian Lake still poses some questions (Gábris et al., 1986) in case of the Maros River system as well (Mike, 1991).

Climatic factors, such as mean annual temperature and annual precipitation have an influence on channel pattern by determining runoff the type of weathering and vegetation cover (Schumm, 1985; Mackey, 1993; Bridge, 2003). In the temperate zone the dimensions of a fluvial system is usually adjusted to annually or biannually returning floods, however under changing climatic conditions the role of single, extreme events may be much more significant (Schumm, 1985; Bridge, 2003). Nevertheless, in terms of alluvial rivers the role of bankfull discharge is stressed as channel forming processes are related mostly to these in various works (Leopold and Wolman, 1957; Richards, 1982; Schumm, 1985; Bridge, 2003). In the Pannonian Basin, located in a climatic transition zone, variations could be significant in the past and could result tremendous changes in the dimensions and capacity of lowland rivers (Gabris, 1986). However, climate change and tectonic activity can be equally apparent and it is particularly difficult to know if a sedimentological and morphological change is due to which of the above controlling factors. (Krzyszkowski, 1996; Bridge, 2003; Vanderberghe, 2003).

Recent activity in data collection on present day rivers has allowed on many parts of the world to develop equations which can be applied to reconstruct past discharge values (Bridge, 2003). The relationship between discharge and planform parameters is well demonstrated in case of meandering rivers (Williams, 1984; Bridge, 2003). Parameters mostly used for the reconstruction are chord length, radius of meander and wavelength of meandering (Williams, 1984). These parameters are sometimes not easy to assess, due to the low number of detectable palaeo-channels, or blurred morphology. Discharge calculations of this type obviously will not work on straight or braided rivers.

Another approach for determining discharge is the application of hydraulic parameters, such as area of cross-section, slope, grain size and roughness. Relationships of this type are more or less based on the classical equation determined by Gauckler and Manning (1890s). The Gauckler-Manning equation has been reworked by several authors to fit to different channel patterns and different sets of empirical data (Lane, 1957; Leopold and Wolman, 1957; Osterkamp, 1978; Begin, 1981), however the original equation is still widely used in the literature (Bridge, 2003). When applying hydraulic parameters for the discharge calculations of palaeo-channels it is of major concern to determine values of original slope and cross-sectional area. In these investigations the role of topographical and sedimentological measurements is inevitable, however shallow geophysical methods, such as ground penetrating radar (GPR), or electrical resistance tomography (ERT) can be also efficient if environmental conditions are suitable (Gourrya et al., 2003; Carreon-Freyer et al., 2003; Bersezio et al., 2007; Van Dam, 2010; Yadav et al., 2010; Rucker et al., 2011)

Considering the above, the first aim of our study was to test morphological and hydraulical equations on a braided and meandering palaeo-channel section located on the alluvial fan of the Maros River in Hungary. For the calculations planimetric, topographic, sedimentological and geophysical data were applied. Henceforth the potential of different geophysical methods could be assessed. Finally, we aimed to calculate an average bankfull, or channel forming discharge, characterising the investigated palaeo-channel reach.

STUDY AREA

The study area is located on the interfluve area bordered by the Tisza, Maros and Körös rivers. Geomorphologically, the territory is primarily dominated by the alluvial fan of the Maros River (Pécsi, 1959). The Maros Alluvial Fan (MAF) with a total area of 10 000 km^2 hosts numerous palaeo-channels, representing Late Glacial, Holocene river generations of the Maros (Borsy, 1989; Molnár, 2007; Sipos et al., 2012). From these, one of the largest palaeo-channel is located in the axis of the alluvial fan, near its western edge at the town of Orosháza (*Fig. 1*).

The present day catchment of the river is 30 000 km^2, most of which is located in the Transylvanian Basin, the Eastern and Southern Carpathians (Molnár, 2007). There are no signs for significant changes in the size of the upland catchment during the Quaternary (Mike, 1991). On its lowland reach the Maros has built a large alluvial fan during the Late Pliocene and Pleistocene with a radius of 80-100 km starting from the Lipova gorge. Changes in the direction of its flow on the fan were caused partly by tectonic (local uplifts and subsidences) and geomorphological (avulsion) factors (Borsy, 1989). The apex of the fan is at 130 m asl while its rim is between 80-85 m asl. The slope of the fan is not uniform, usually it is between 20 and 30 cm/km however near the edge there is a belt where it can reach occasionally 50 cm/km (Sümeghy and Kiss, 2011).

Thanked to the above, the river has a relatively high energy which is also resembled by the high sediment transport capacity of the present-day Maros, which is similar to that of the Tisza or the Danube. Horizontally, the sediment of the MAF changes from primarily gravel near the apex to sand and silt towards the edges (Borsy, 1989). Vertical variations in the granulometry of the fan sediment are related to Pleistocene climatic changes in general, namely coarser and finer strata referring to glacial and interglacial phases, respectively (Mike, 1991; Molnár, 2007).

The Hungarian part of the MAF can be divided into four geomorphological subunits based on surface forms and sediments (Pécsi, 1959). The central part is covered by sandy loess and dominated by meandering palaeo-channels (*Fig. 1*). The NE wing is mainly composed of alluvial loess and silt deposits and hosts both meandering and braided palaeo-channels. The northern edge just south of the Körös Basin is covered with sandy loess and primarily braided channel forms are characteristic. Finally the western wing has alluvial loess on the surface and meandering channels.

The site of the investigation at Orosháza is right at the edge of the alluvial fan. East of the town the almost 1 km wide channel has a braided pattern with several sub channels and huge bar forms. Downstream the same channel after leaving the edge of the fan becomes meandering, its width decreases at certain sections to only a few 100 m. According to results of OSL dating the Orosháza channel was active during the Pleistocene-Holocene shift, between 14 and 8 ka (Sipos et al., 2012). As a consequence, post formational loess deposition could not be significant. Partly because of this and due to supposedly fast avulsion processes (Sümeghy and Kiss, 2011) forms are still sharp and well detectable. For the present research two study sites were investigated, one on the braided, the other on the meandering section of the palaeo-channel (*Fig. 1*).

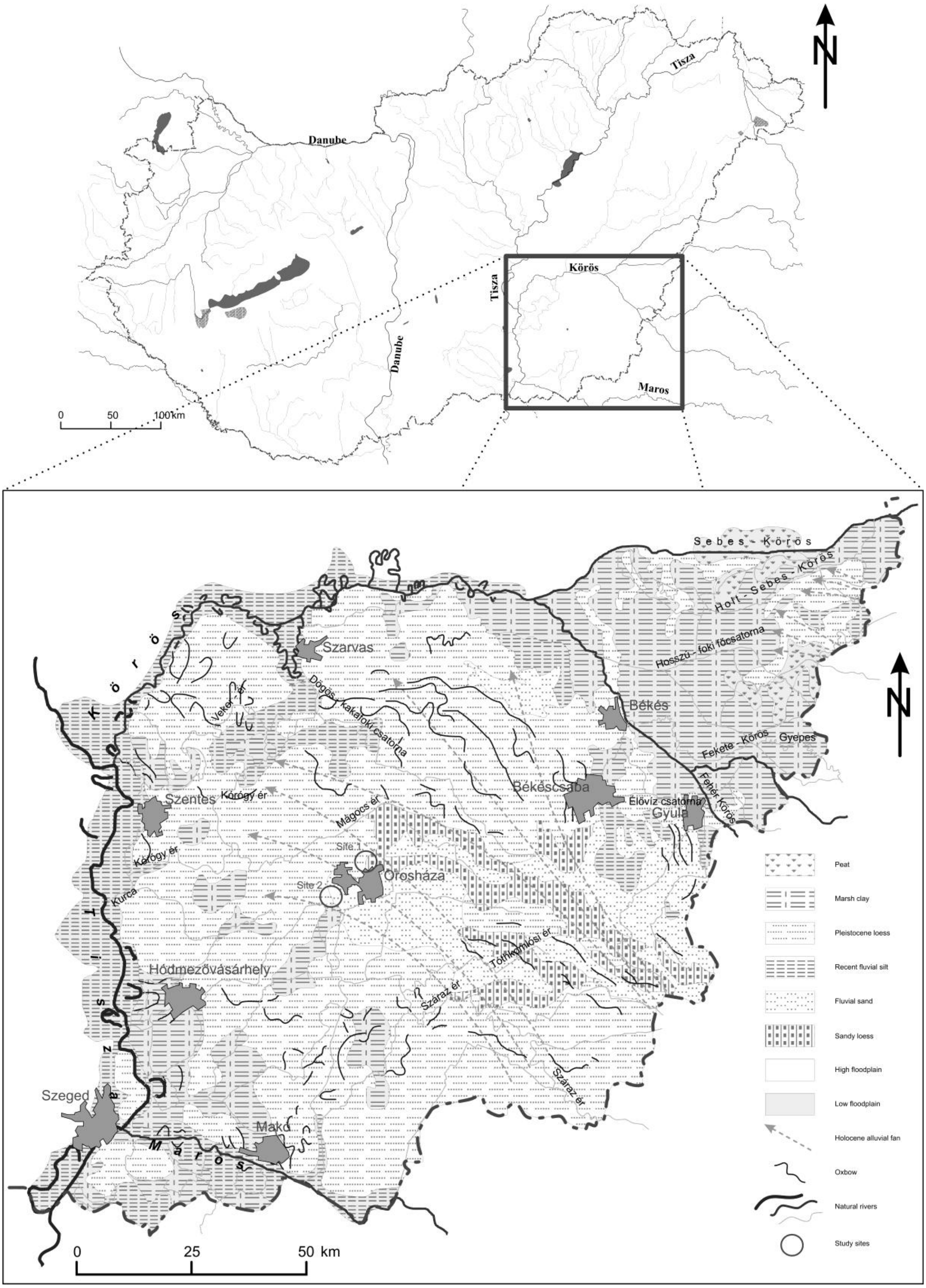

Fig. 1 Location of the study area and the Geomorphological Map of the region (based on Pécsi, 2000)

METHODS

As a first step a digital elevation model (DEM) was produced from 1:10000 scale topographic maps with 1 m interval primary contour lines, supplemented by 0.5 m interval secondary contour lines on flat areas. The DEM was made with the ArcGIS Topo to Raster interpolation tool. Vertical resolution was better than 1 m, while horizontal raster resolution was 10 m (*Fig. 2*). The DEM was used to determine channel planimetric parameters (meander wavelength, meander radius), present day silted up cross-sectional parameters (width, deth, area), and channel slope conditions on the investigated palaeo-reach. Cross-sectional indicators were measured at various points in order to get average parameters. The area's general slope was calculated on the basis of SRTM data.

Since the DEM enabled only the calculation of modified, silted up cross-sectional parameters other methods also had to be applied and compared to determine the original dimensions of the river. The primary aim was to determine the depth of channel sediments and calculate the true depth of the river. In all 16 drillings were made in two cross-sections representing both the braided and the meandering reaches (*Fig. 2*). Drillings were representing characteristic topographic features, such as the natural levee, river bed, bar crests, point bars and chutes. Bore holes were deepened till the level of the groundwater. The depth of the drillings in the channel was 1.5-2 m, on point bars and levees 5-6 m. Samples were taken at every 10 cm. Sedimentological analysis included grainsize analysis. The grainsize-distribution of the samples was determined by a Fritsch Analysette 22 laser equipment with a measurement range of 0.08-2000 μm. Samples underwent ultrasonic homogenisation and all measurements were repeated three times to check if there is further disintegration. For the geomorphological interpretation sample D50 values were determined.

In order to test the applicability of geophysical methods in determining sedimentary shifts GPR and ERT were applied in the drilling sections (*Fig. 3*). GPR measurements were made with a GSSI type instrument, using 200 and 270 MHz shielded antennae. Unfortunately right during the first field work it turned out that sedimentary conditions (high clay and silt content in the upper strata) disable the successful use of GPR technology on these sites. Therefore, ERT profiling was favoured during later measurements. For these measurements a PASI type 32 electrode system was used. ERT sections were measured using a Wenner electrode array. Electrode spacing was 5 m, one section

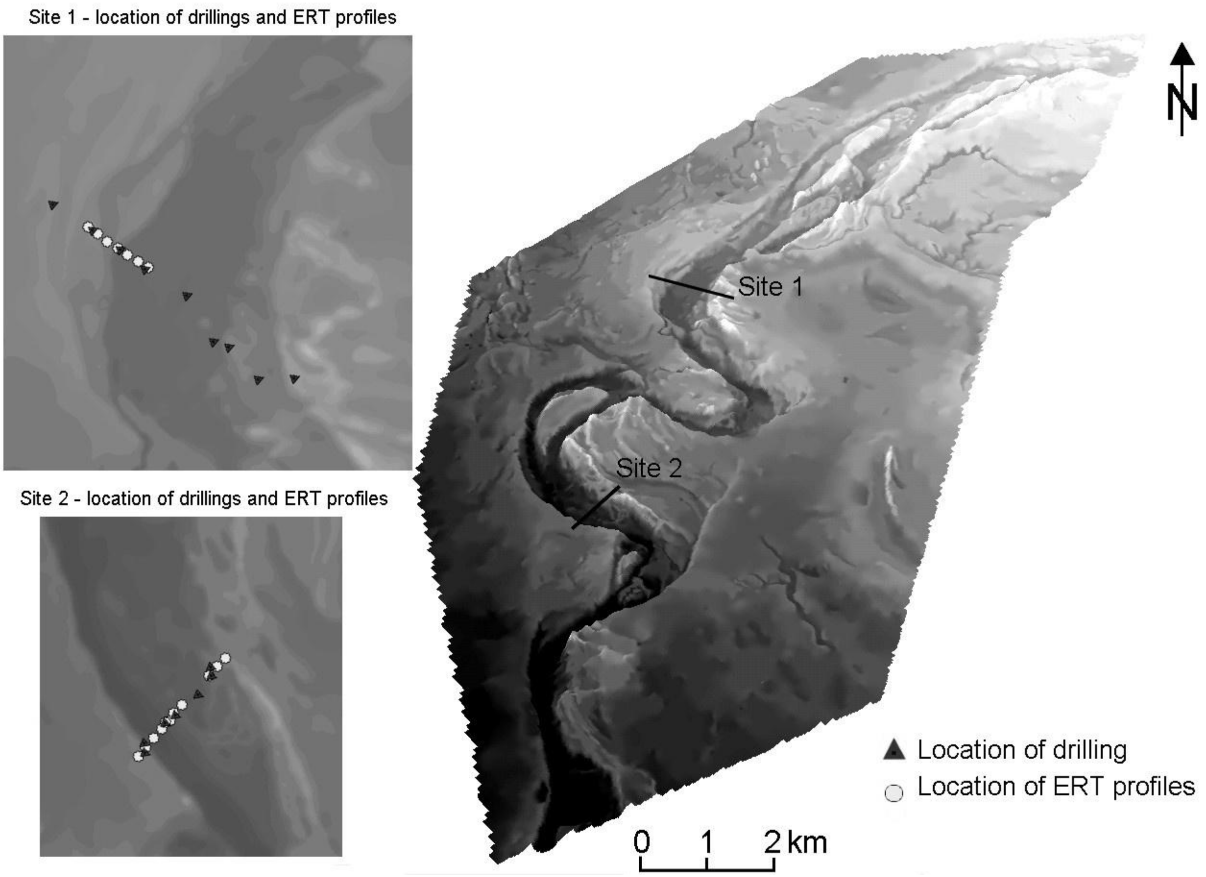

Fig. 2 DEM generated from 1:10000 scale topographic maps with the cross-sections studied in detail

Fig. 3 Geophysical field investigations. The bank of the braided section on the left, and bank of meandering on the right.

was therefore 160 m, from these 5 were measured on both sites. Neighbouring sections had a 50 % overlap. Maximum penetration was 20-30 m, data were collected from 10 levels.

From a geomorphological aspect bankfull discharge is one of the most important parameter of river flows, as it is highly responsible for the geometry of the cross-sections and the channel pattern as well (Schumm, 1985). Bankfull discharge was assessed by selecting and using equations from the literature, with special attention to the range of applicability of the published formulae.

In terms of planimetric parameters the equations of Leopold and Wolman (1957) and Mackey (1993) were selected (*Table 1*). These formulae can be applied only for meandering channels, and they use meander wavelength as the base parameter. The relevance and significance of meander wavelength was also reinforced by Timár and Gábris (2008) when studying the channels of the Great Hungarian Plain. The selected formulae operate up till the 1000 m^3/s discharge range, thus proved to be adequate as a first approximation.

Discharge calculations were also made from cross-sectional parameters. At the two measurement sites these were derived from DEM (present) and sedimentological (original) analyses, results were compared and coefficients were calculated. These coefficients were used to calculate average original cross-section parameters from representative present-day values calculated by taking the mean and standard deviation of several measurements on the DEM.

In case of the meandering section the relationships determined by Dury (1976) and Williams (1978) were tested. The range of the applicability of the Williams (1978) formula ($0.5<Q_b<28320 m^3/s$; $0.7<A_b<8510 m^2$; $0,000041<s<0,081$) is well over the expected discharge of the investigated system (*Table 1*). In terms of the braided reach the modified Gauckler-Manning equation was applied. The basic equation ($Q=AR^{2/3}S^{1/2}/n$) was modified and reduced by Rotnicki (1983) based on more than 1000 observations on rivers with highly variable geometry ($Q=(1,49/n)wd^{5/3}s_c^{1/2}$). Parameter “n” was determined after Goodwill and Sleigh (2002), and calculations made on the basis of present-day measurements on River Maros (Katona et al., 2012). The average value used for calculations was 0.056.

RESULTS

Sedimentological and geomorphological interpretation

The original channel cross-sections were reconstructed on the basis of sedimentological data and geophysical measurements. Channel sediments can be clearly identified in the sedimentary columns, thus the amount of silting up in the thalweg zone can be estimated to be approximately 2-3 m (*Fig. 4*). Silting up is suggested to be the result of post-formational fluvial activity as from time to time the channel could be partially reactivated during the extreme floods of the

Table 1 Equations applied for the calculation of bankfull discharge

meandering	author	equation
planform	Leopold and Wolman (1957)	$L=65.2Q_b^{0,5}$
	Mackey (1993)	$L=72.16Q_b^{0,49}$
	Sümeghy and Kiss (2011)	$Q = 0.0003*L^2+0.3440*L-81.329$
cross-section	Dury (1976)	$Q_b=9.93 A_b^{0,85}$
	Williams (1978)	$Q_b=4.0A_b^{1,21}s_c^{0,28}$
braided	author	equation
cross-section	Grauckler-Manning	$Q=wd^{5/3}s_c^{1/2}*1.49/n$

Maros system. Natural levees, point bars and mid channel bars are still sharply detectable. We took the present level of these as the minimum heights of banks as subsequent erosion and agricultural activity could lower their level. The D50 value of overbank, or suspended sediments is similar at the two study sites, being 25-30 and 20-25 μm in the braided and the meandering corss-section, respectively (*Fig. 4*). Nevertheless, there is an abrupt change in the grainsize of the bedload. On the braided section the D50 value of channel sediments is 200-300 μm, representing the range of coarse grain sand. Meanwhile, less than 15 km downstream on the meandering section, mean grainsize decrease to 50-150 μm. The degree of sorting however also decreases, but this is due to the appearance of clayey fractions in channel sediments.

In terms of channel slope and general slope it turns obvious that the upstream braided section has a 15-20 % lower slope than the downstream meandering one (*Table 2*). The knick point signs the transition between channel patterns. Such change in slope is unusual in terms of normal gradient alluvial rivers (e.g. Schumm, 1985; Richards, 1996). In this case a possible cause could be tectonic activity, however, this does not explain the significant change in the composition of channel sediments. Pattern and slope change is rather caused by the geomorphological background, i.e. the knick point marks the edge of the alluvial fan.

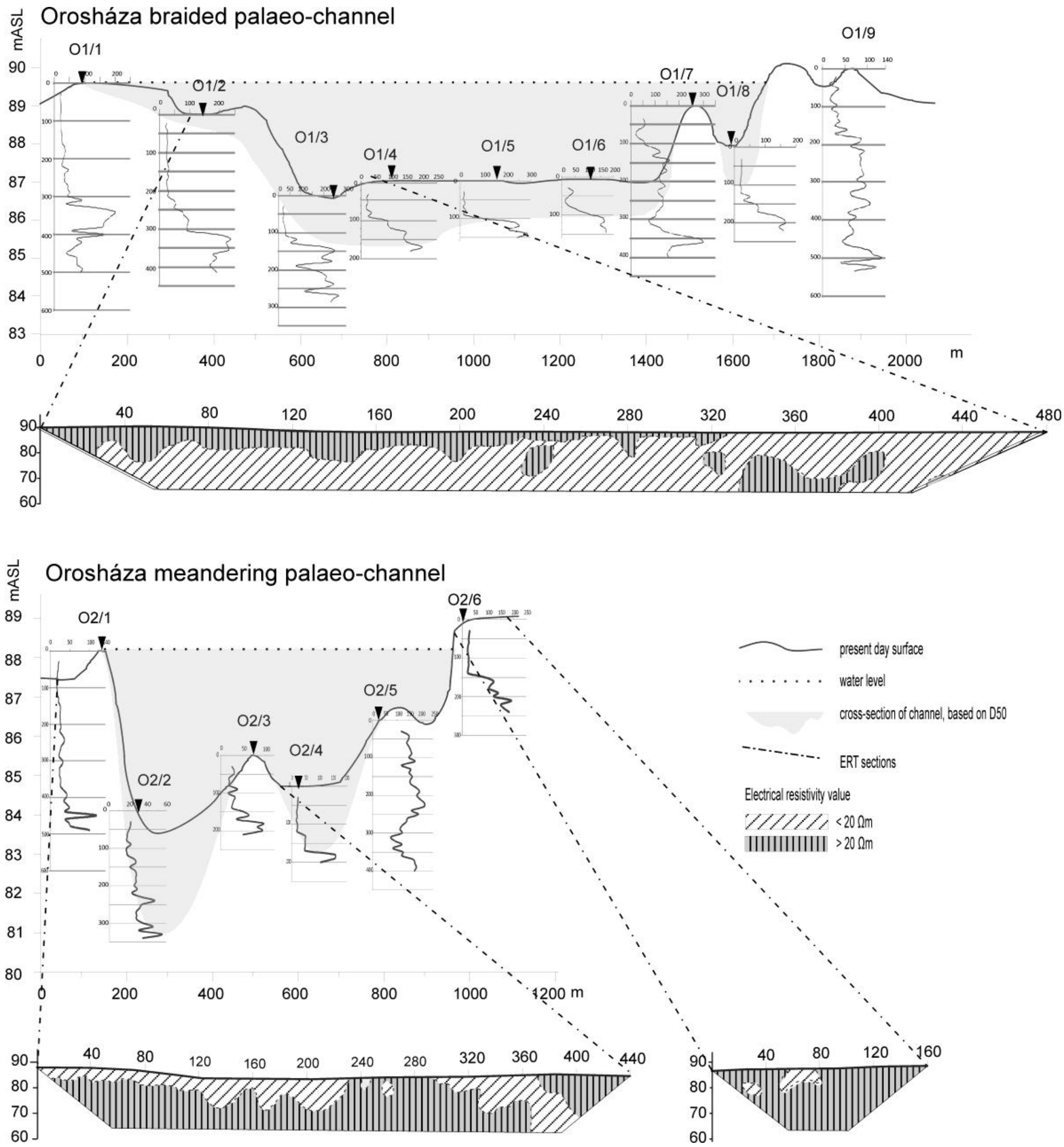

Fig. 4 Cross-section of the sites with sedimentary sequences of drillings and generalised ERT profiles

By reaching the distal parts of the fan the competence and capacity of the river, having a very high bed load of coarse sand, decreased considerably, leading to the development of an aggradation zone (braided section). Downstream of this zone slope and energy increased, and as the river already deposited its coarse sediment upstream, meandering was favoured. Meandering could be further facilitated by sedimentological differences, i.e. in front of the coarser alluvial fan sediments finer more cohesive floodplain deposits are located, confining the banks of the channel (*Fig. 1 and 4*).

Difference in slope suggests a difference in cross-sections as well, namely, larger slope means smaller cross-section at similar discharge on the same river. Data have reinforced this relationship (*Table 2*). The average cross-section of the braided and meandering reaches could be around 3000 and 3800 m^2, respectively. Width/depth ratios also show a remarkable difference between the two sections, being around 400 on the braided and 130 on the meandering section.

Geophysical interpretation

In an alluvial setting the electric resistance threshold between sedimentary units is hard to determine, as different sediments can have very similar resistance (Reynolds, 1997). In general however, the coarser and drier the sediment the higher resistance is received. The study area is primarily characterised by mixed silt, fine and medium sand, which made the interpretation of ERT profiles complicated.

The ERT measurements were made in autumn among dry conditions, thus precipitation could not influence resistivity parameters near the surface (Reynolds, 1997). The received values were between 0 and 300 Ωm. As a first approximation the threshold between silt and sandy sediments was based on the values reported by Nádor et al. (2005) and Yadav et al. 2010, and was taken 20 Ωm. Resistivity profiles however could not entirely be compared to the drilling profiles, due to the resolution provided by the 5m electrode spacing. Nevertheless, in case of Site 1 (*Fig. 5*) the top layers, especially near the bank of the channel, were dominated by sediments with high resistance down till 10-15 m (75-80 m asl). Note that ground water appeared at around 85 m asl, therefore the change experienced in this case is primarily due to changes in grainsize, meaning that coarser sandy sediments are deposited on finer clayey deposits. High resistivity strata wedge out towards the centre of the channel.

At Site 2 resistivity values were a little higher in general (*Fig. 6*), which seems to contradict the hypothesis that sedimentary deposits are finer in front of the alluvial fan. Contradiction can only be relieved if we suggest that the mineral composition of sediments at the two sites is slightly different. This hypotheses however needs further testing. Nevertheless, relative differences of the Site 2 profile show an opposite pattern than that of Site 1 profile. Lower than 20 Ωm values appear on the top left, while higher resistivity values characterise sediments 5-10 m below the surface. As the boundary is again beneath the ground water table it is obvious that palaeo-channel sediments were laid on coarser deposits. Lateral bar surfaces in the middle of the cross section have higher resistivity similar to point bar sediments on the right (*Fig. 6*). Based on these data, the boundary surface might represent the base of the palaeo-channel, although the 20 Ωm threshold needs to be revised in light of higher resolution ERT results. In all, further calculations were based on the results yielded by sedimentological analysis, at present resistivity profiles are not precise enough to draw unambiguous sedimentary boundaries.

Palaeo-discharge calculations

Bankfull discharges calculated by different approaches showed considerable variance (*Table 3*). In terms of the meandering reach discharges calculated on the basis of

Table 2 Cross-sectional parameters of the investigated sites

	Site 1 (braided)			**Site 2 (meandering)**		
L (m)	-			2500 ± 600		
s (cm/km)	25.6 ±1.3			29.7 ± 1.0		
s_c (cm/km)	20.3 ± 0.3			22.5 ± 1.0		
	Parameters of the investigated cross-sections					
	DEM	drilling	coeff.	DEM	drilling	coeff.
w (m)	1125	1125	1,00	685	685	1.00
d (m)	2,3	2,8	1,21	3,15	4,20	1.33
A_b (m^2)	2345	3150	1,34	2076	2870	1.38
	Average values based on several measurements on the DEM					
	DEM	original		DEM	original	
average w (m)	1120 ± 140	1120 ± 140		595±80	595±80	
average d (m)	2,2±0,32	2,7 ± 0,7		3,05±0,7	4,05±0,9	
average A_b (m^2)	2465±470	3025 ± 870		1815±480	2410±625	

L: meander wavelength, s: general slope, s_c: channel slope, w: width, d: deep, A_b: area of bankfull cross section

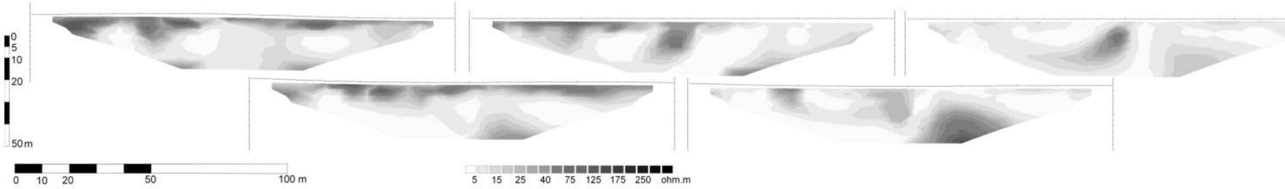

Fig. 5. ERT profiles at Site 1

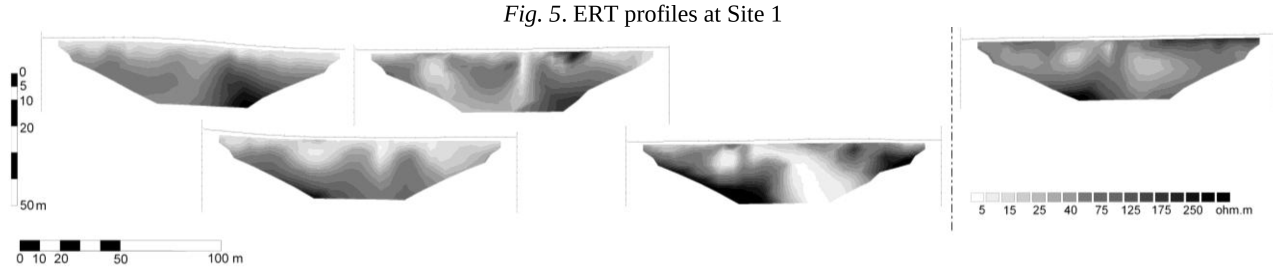

Fig. 6. ERT profiles at Site 2

the classical planform equations (Leopold and Wolman, 1957; Mackey 1978) were considerably, almost 50 % lower than the value received using the region-specific formula of Sümeghy and Kiss (2011). By using the equation of Dury (1976), setting up a relationship between discharge and cross-sectional area in case of meandering rivers, the received result was clearly and significantly overshooting the previous values. When slope was introduced to the relationship (Williams, 1978) the estimated discharge decreased, but it was still significantly higher than the values derived from planform parameters. By applying the modified Gauckler-Manning equation the calculated discharge was very close to the results yielded by the Sümeghy and Kiss formula. Based on this, we assume that the two approaches reinforce each other and the bankfull discharge of the river on the meandering section was approximately 2500 m^3/s.

In case of the braided section the only way to determine bankfull discharge was to use the Gauckler-Manning equation. In theory the result received for this section should be similar to that calculated for the meandering reach. As a matter of fact, results were fairly well corresponding, fell within the error limits and there was only a 9 % difference between the mean values. Consequently, the Gauckler-Manning equation seems to be suitable to determine bankfull discharge values on braided channel section on the MAF. It has been proved however that classical planform formulae significantly underestimate, while cross-section based formulae overestimate the discharges in this environment. One reason can be, that based on width/depth ratios the channels – even those meandering – of the MAF are transitional.

If we compare the received values to the present-day bankfull discharge (850 m^3/s) of the Maros River (Fiala et al., 2007), there is a considerable difference, referring to a much different flow regime during the onset of the Holocene.

Table 3 Bankfull discharge values calculated from planform and original cross-sectional parameters.

Equation	Site 1	Site 2
Leopold and Wolman (1957)		1470 ± 350 m^3/s
Mackey (1993)		1390 ± 330 m^3/s
Sümeghy and Kiss (2011)		**2650 ± 630 m^3/s**
Dury (1976)		7280 ± 1980 m^3/s
Williams (1978)		4570 ± 1240 m^3/s
Gauckler - Manning	**2220 ± 640 m^3/s**	**2445 ± 645 m^3/s**

CONCLUSIONS

The Orosháza palaeo-channel system provided a good opportunity to test different field and computational methods for determining the bankfull discharge of meandering and braided channel sections. The complex analysis included geomorphological, geophysical and sedimentological methods, and emphasizes the necessity of comparing these results.

Based on the investigations some important geomorphological results could be drawn. First, a discrepancy between slope and channel pattern was experienced on the area, which is probably due to inherited slope conditions and high bed load deposition near the edge of the MAF. This suggests that as a matter of an avulsion event the studied channel was not operating as a main flow path for a long time, i.e. channel slopes could not be adjusted to different general slope values. Since their formation the channel silted up 2-3 m, meaning that the forms studied are fairly young, thus the effect of post formational tectonic changes might be also of secondary importance.

At the present phase of research sedimentological and geophysical profiles could not entirely be overlapped, as a matter of their different vertical extent and resolution. However, ERT profiling seems to have a great potential in the fast detection of sedimentary structures on the study area if electrode spacing is decreased. GPR measurements failed due to the high clay and silt content of the sediments.

Discharge calculations on the meandering and braided reaches showed an acceptable agreement. In case of the meandering reach the results based on the meander wavelength/bankfull discharge formula of Sümeghy and Kiss (2011) and those derived from the Gauckler-Manning equation reinforced each other. Moreover, the value received for the braided reach was also in a good correspondence, meaning that cross-sectional parameters determined from sedimentological results can have an important role in estimating the bankfull discharge of braided channels on the MAF.

In all, the calculated bankfull discharges suggest much higher energy fluvial processes than today. Considering the age and the dimensions of the Orosháza palaeo-channel, and also taking into account that the area of the Maros catchment has not changed in the past 10-15 ka, we can suggest that high discharges were mainly related to the melting of glaciers in the upland drainage basin.

Acknowledgements

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