Fluvial response to proglacial effects and climate in the upper Dnieper valley (Western Russia) during the late Weichselian and the Holocene

Réponse du Dniepr supérieur (Russie occidentale) aux influences proglaciaires et au climat au Weichselien supérieur et à l’holocène

Andrei Panin, Grzegorz Adamiec and Vladimir Filippov
ABSTRACT

The Dnieper River in the westerly-aligned reach upstream from the Russian/Belarusian border exhibits deep early MIS 2 (pre-LGM) incision that is considered to have been caused by valley tilting associated with the growth of glacial forebulge. Glacial damming occurred around the LGM downstream from the city of Smolensk, locally shifting the valley southwards and stimulating substantial aggradation, although in a limited reach of few tens of kilometres. The failure of the glacial dam induced incision down to pre-LGM elevations, which was followed, before the onset of the Holocene, by further aggradation that was probably a response to a decrease in valley gradient caused by the subsidence of the glacial forebulge. This deformation continued into the Holocene and is estimated to have caused valley tilting by 4-5 m within a 100 km valley reach, which is noticeably larger than that predicted by the ICE-5G (VM2) model (Peltier, 2004) and suggests a delayed forebulge collapse in comparison with that predicted by modeling. In the Early Holocene, the river formed very wide braided or single-thread palaeochannels. Early Holocene channel morphology implies discharges estimated at three times that of the present-day river, which indicates a considerable increase of discharge from the Upper Dnieper catchment, presumed to be climatically driven. Substantial remnants of Moscovian (MIS 6) glacigenic deposits, preserved as intra-valley residual hills, are interpreted as having been dissected and isolated during valley initiation at the end of the Moscovian and then partly buried by fluvial sediments due to valley aggradation prior to the onset of the Holocene.

Keywords: Dnieper, glacial forebulge, glacial-dammed lake, large palaeochannel, palaeodischarge

FLUVIAL RESPONSE TO PROGLACIAL EFFECTS AND CLIMATE IN THE UPPER DNIEPER VALLEY (WESTERN RUSSIA) DURING THE LATE WEICHSELIAN AND THE HOLOCENE

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1 INTRODUCTION

Large palaeochannels are known from many regions of Central and Eastern Europe (Szumanski, 1983, 1986; Vandenberghhe et al., 1994; Howard et al., 2004; Borisova et al., 2006; Gębica et al., 2009; Sidorchuk et al., 2009; Kasse et al., 2010). In most cases they have a meandering pattern and a Lateglacial age. Most researchers associate large meanders with significant climatically-driven increases in fluvial discharge. Large palaeochannels ten or more times the width of the modern river were also recognized in the upper and middle Dnieper valley. Unlike in other European valleys, where large Lateglacial meanders form part of low terraces or the floodplain, many large palaeochannels in the Dnieper valley are separated from the present-day river by hills composed of Mid-
Pleistocene till (Barashkova et al., 1998). Such palaeochannels were studied between Orsha and Rogachev in Belarus and assigned a LGM age (MIS 2) on the basis of geomorphic correlation with terraces in tributary valleys (Kalicki & San’ko, 1992, 1998). Large palaeochannels were interpreted as branches of a large glacially-fed river that were abandoned after the glacial meltwater inflow had ceased.

The role of the Dnieper fluvial system in the southward transportation of glacial meltwater cannot be overstated. The Dnieper River was the only route linking the Black Sea to the decaying Scandinavian ice sheet (SIS; fig. 1) in late MIS 2. The occurrence of glacial meltwater pulses has been widely used to interpret the Black Sea hydrology during the Late Pleniglacial (Major et al., 2006; Lericolais et al., 2011; Sidorchuk et al., 2011; Soulet et al., 2011, 2013). Despite the fact that glacial meltwater outflow through Dnieper and its tributaries has been important in the interpretation of Black Sea and Caspian Sea palaeogeography during the period of deglaciation, the timing and path of this outflow remains unsure. Also neglected has been the role in valley development of crustal warping related to glacio-isostatic effects, which may be deduced from the close location of the Upper Dnieper valley to the SIS boundary (fig. 1 and 2). It has long been recognized that horizontal mass transfer in the low viscosity asthenosphere due to glacial loading would have induced uplift and the formation of a peripheral bulge with its axis parallel to the ice sheet boundary (e.g. Mörner, 1979; Peltier, 1990). Estimates of forebulge axis uplift at the south-eastern margin of the SIS range from 60 m (Fjeldskaar et al., 2000) to as much as 110 m (Peltier, 2004) and 170 m (Mörner, 1979). Crustal unloading due to ice sheet retreat was accompanied by forebulge subsidence. The influence of proglacial crustal movements on fluvial systems at the periphery of the SIS was studied by Bylinski (1990, 1996), who described changes of alluvial terrace thickness in river systems from the Elbe in the west (Germany) to the Pechora in the north-east (Russia). These were related to river damming by the forebulge, and erosion/aggradation successions associated with forebulge migration in response to a retreating ice margin during deglaciation. Wallinga et al. (2004) and Busschers et al. (2007) found incision/aggradation events and lateral valley tilting in the Rhine-Meuse delta around the LGM and related them to the build-up and subsequent subsidence of a SIS glacio-isostatic forebulge. The position of the Upper Dnieper in the vicinity of the ice margin (fig. 1 and 2) makes it potentially vulnerable to similar effects associated with SIS forebulge rise and decay.

The aim of the present study is to use a multi-proxy approach, including field-survey, numerical dating and geomorphological correlation, to reconstruct the development of the Upper Dnieper during the last 20-25 ka, including the origin and chronology of large palaeochannels, and to assess the respective roles of proglacial effects, such as forebulge formation and glacial damming, as well as palaeohydrological change, on this development.
2 - STUDY AREA

The upper Dnieper River valley was most probably formed at the end of MIS 6 after deglaciation of the Moscovian (Late Saalian) ice sheet, although it has been hypothesized (Kvasov, 1975, 1979) that this reach of the valley was formed only in post-LGM time due to drainage of large proglacial lakes. The present-day Upper Dnieper is a medium-sized river with an annual discharge of 43.5 m³.s⁻¹ at Dorogobuzh (catchment area 6,390 km²) and 98.7 m³.s⁻¹ at Smolensk (catchment area 14,100 km²) (fig. 2). The hydrological regime of the modern river is characterized by the highest stage occurring during the regular snowmelt flood (typically in April) and a low water stage persisting throughout the rest of the year, which may be interrupted by occasional rain-fed floods. The mean maximum (spring flood) discharge is 690 m³.s⁻¹ at Dorogobuzh and 830 m³.s⁻¹ at Smolensk. Rain-fed floods in summer-autumn and thaw floods in winter are always much lower than the spring flood. The river channel is 80-100 m wide at bankfull stage. The channel has an irregular pattern: long straight stretches alternate with short series of 2-3 meander bends with varying curvature. The long profile is concave upstream from the study area, but nearly straight at the 150 km stretch between Dorogobuzh and Smolensk (fig. 3).

The Upper Dnieper catchment is a combination of hilly uplands with a maximum elevation of 260-280 m a.s.l., and flat swampy lowlands lying at 170-200 m a.s.l, the Dnieper being located at 175 m a.s.l. in Dorogobuzh and 160 m a.s.l. in Chekulino. Bedrock is everywhere covered by Quaternary sediments deposited mainly during Middle and Late Pleistocene glaciations. In the Late Valdai (Late Weichselian, MIS 2), glacial melt waters probably joined the Dnieper valley through its right-bank tributaries such as the Vop’, Hmost’, and Berezina (Barashkova et
Fig. 3: Upper Dnieper valley long profile and elevation of palaeochannels, modern and estimated for the Early Holocene.
1/ Valley long profile, 2/ valley gradient, 3/ river sinuosity; Early Holocene palaeochannels (top and base of channel deposits), 4/ modern position in the valley (based on sections 4.2 and 4.4 and figures 6 and 8), 5/ estimated for the Early Holocene (see sections 4.5 and 4.6); vertical arrows show amplitude of valley tilt.

Fig. 3 : Profil en long de la vallée du Dniepr avec l’altitude des chenaux modernes et début-Holocène. 1/ Profil en long de la vallée, 2/ pente de la vallée, 3/ sinuosité du fleuve; paléochenaux du début de l’Holocène (sommet et base des alluvions), 4/ position actuelle dans la vallée (cf. sections 4.2 et 4.4 et figures 6 et 8), 5/ position estimée pour le début de l’Holocène (cf. sections 4.5 et 4.6); les flèches verticales indiquent l’ampleur du changement de pente de la vallée.

Fig. 4: Glacial forebulge at the SE margin of the SIS at 21 ka, predicted by the ICE-5G (VM2) model (Peltier, 2004).
Contour lines show topography at 21 ka minus topography at 0 ka. Arrows point to locations of study sites (1/ Chekulino, 2/ Gnezdovo, 3/ Solvyovo).
The ice-covered area is hatched.

Fig. 4 : Bourrelet périphérique en marge SE de l’inlandsis scandinave il y a 21 ka, prédit par le modèle ICE-5G (VM2) de Peltier (2004). Les courbes de niveaux indiquent la différence entre la topographie il y a 21 000 ans et la topographie actuelle. Les flèches indiquent la localisation des sites d’étude (1/ Chekulino, 2/ Gnezdovo, 3/ Solvyovo). L’espace occupé par l’inlandsis est hachuré.
The influence of the SIS on valley development also included crustal warping over the proglacial area sufficient to modify valley gradient. According to the current version of the ICE-5G glacio-hydro isostasy model (Peltier, 2004), the upper Dnieper would have been located on the proximal side of the LGM peripheral forebulge (fig. 4). At the sub-latitudinal valley section between Dorogobuzh and Orsha, crustal updoming would have led to increased valley gradient during SIS advance and its subsequent decrease due to forebulge collapse as a result of deglaciation.

3 - METHODS

3.1 - GEOMORPHOLOGICAL SURVEY

To identify landforms in the Dnieper valley we used space images with 1-15 m resolution (Landsat 7 panchromatic channel, Google Earth, Yandex Map) and topographic maps with scales between 1:25,000 and 1:100,000. Even the most detailed maps with contour interval of 5 m were insufficiently precise to distinguish between different elements of relatively flat floodplain surface. However, a combination of maps with images obtained during spring flood and in autumn allowed a better understanding of the floodplain topography and measurement of palaeochannel parameters. In areas of swale-and-ridge topography created by lateral migration of meandering channels, palaeochannel width can be roughly estimated from the “two-thirds rule” adopted in fluvial sedimentology (Ethridge & Schumm, 1978), which derives from Allen’s (1965b) statement that point bars occupy two-thirds of a channel width.

Analysis of satellite images was complemented with field recognition of landforms. Topographic profiling and precise measurement of landform elevation above the river were conducted at study sites by DGPS field survey with a Leica Smart Station. Conversion from ellipsoid WGS84 heights to orthometric heights (leveled elevations above mean sea level) used on topographic maps was made with the EGM2008 geoid model. Sedimentary composition of valley forms was studied in hand and mechanical cores (up to 22 m deep), in hand pits and in river-bank exposures.

3.2 - NUMERICAL DATING

An absolute chronology was established by radiocarbon and OSL dating. Samples were taken from vertical cleaned sections or from cores. OSL dates were produced in the GADAM Centre of Excellence, Institute of Physics, Silesian University of Technology, Poland (Panin et al., 2014; tab. 1). For dose rate determination,

<table>
<thead>
<tr>
<th>No.</th>
<th>Core / section name</th>
<th>D, m</th>
<th>Dm, m</th>
<th>WC, %</th>
<th>DR, Gy/ka</th>
<th>ED, Gy</th>
<th>NA</th>
<th>Age, years¹</th>
<th>Index GdTL</th>
<th>Source²</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Ch-11-02</td>
<td>3.0-3.5</td>
<td>3.3</td>
<td>18</td>
<td>1.826±0.068</td>
<td>13.20±0.10</td>
<td>19/10</td>
<td>7160±620</td>
<td>1465</td>
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</tr>
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<td>1.4</td>
<td>1.4</td>
<td>18</td>
<td>1.591±0.057</td>
<td>13.93±0.75</td>
<td>23/14</td>
<td>8690±570</td>
<td>1466</td>
<td>1</td>
</tr>
<tr>
<td>3</td>
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<td>1.6-2.0</td>
<td>1.8</td>
<td>30</td>
<td>1.735±0.067</td>
<td>15.77±0.61</td>
<td>22/22</td>
<td>9090±510</td>
<td>1467</td>
<td>1</td>
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<td>Gn-10-02</td>
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<td>8.0</td>
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<td>13/11</td>
<td>21400±2800</td>
<td>1238</td>
<td>1</td>
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<td>10.0</td>
<td>14</td>
<td>2.161±0.091</td>
<td>13.52±0.46</td>
<td>10/10</td>
<td>6200±340</td>
<td>1686</td>
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<td>Gn-11-01</td>
<td>13.5-14.0</td>
<td>12.0</td>
<td>28</td>
<td>2.74±0.13</td>
<td>42.6±2.5</td>
<td>15/15</td>
<td>15500±1200</td>
<td>1687</td>
<td>2</td>
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<tr>
<td>7</td>
<td>Gn-11-03</td>
<td>9.2-9.6</td>
<td>8.0</td>
<td>30</td>
<td>1.478±0.078</td>
<td>7.3±1.1</td>
<td>21/21</td>
<td>4940±760</td>
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<td>1.4</td>
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<td>1.507±0.064</td>
<td>17.55±0.81</td>
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<td>11590±740</td>
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<td>3.5</td>
<td>17</td>
<td>1.890±0.082</td>
<td>102.1±5.6</td>
<td>10/10</td>
<td>54000±3800</td>
<td>1691</td>
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</tr>
<tr>
<td>10</td>
<td>Sol-11-05</td>
<td>1.4</td>
<td>1.4</td>
<td>18</td>
<td>1.718±0.062</td>
<td>17.37±0.96</td>
<td>28/28</td>
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<td>1468</td>
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<td>1.6</td>
<td>1.6</td>
<td>25</td>
<td>2.030±0.076</td>
<td>19.10±0.13</td>
<td>20/20</td>
<td>9410±700</td>
<td>1469</td>
<td>1</td>
</tr>
<tr>
<td>12</td>
<td>Sol-11-02</td>
<td>3.2-3.6</td>
<td>3.4</td>
<td>30</td>
<td>1.433±0.066</td>
<td>17.50±0.11</td>
<td>21/12</td>
<td>12210±950</td>
<td>1470</td>
<td>1</td>
</tr>
</tbody>
</table>

Tab. 1: OSL ages from the studied sites in the Upper Dnieper valley.
Tab 1 : Âges OSL des sites étudiés dans la haute vallée du Dniepr.
the samples were stored for ca. three weeks and high-resolution gamma spectrometry with Canberra HPGe detector was carried out in order to determine the content of U, Th and K in the samples. Each measurement lasted for at least 24 h. The activities of the isotopes present in the sediment were determined against IAEA standards RGU, RGTh, RGK after subtraction of the detector background. Dose rates were calculated using the conversion factors of Adamiec and Aitken (1998). The cosmic ray dose-rate contribution at the site was estimated as described by Prescott and Hutton (1994). Average lifetime of samples relevant for this estimation was assigned with respect to erosion/sedimentation history of each site and may differ from actual depth of samples (tab. 1). The average water content (WC) was assigned on the basis of measurements of saturation water content (SWC) determined using the single-aliquot regenerative-dose (SAR) protocol (Murray and Wintle, 2000), and the age estimates were obtained using the Central Age Model.

Radiocarbon dates relevant for this study (tab. 2) were selected from the geochronological database for the Upper Dnieper catchment (Panin et al., 2014). All dates were produced using liquid scintillation techniques in the Geological Institute of Russian Academy of Sciences (RAS) in Moscow (index GIN), Institute of Geography RAS in Moscow (index IGAN) and Sankt-Petersburg University (index LU). Dates were re-calibrated in OxCal 4.2 (Bronk Ramsey, 2009) with the IntCal13 calibration curve (Reimer et al., 2013).

3.3 - PALAEODISCHARGE AND PALAEOSLOPE ESTIMATION

Flood discharges $Q_T$ that formed palaeochannels can be estimated from mean flow velocities, palaeochannel width $W$ and average depth $D$. Of the three parameters required only $W$ may be measured directly from palaeochannel topography (only palaeochannel cut-offs are suitable as they preserved their former width), the other two needing indirect estimation. Flood flow velocities may be approximated by velocities necessary for bedload transport, or critical flow velocities $V_c$ of sediment entrainment. Maizels (1983) compared five methods of palaeovelocity determination from grain size characteristics of fluvial sediments and found a large scatter of results, which led to the conclusion that each method provides the order of magnitude rather than an accurate value of palaeovelocity. The other problem is that the grain size of fluvial sediments is extremely variable in natural streams, and results of velocity estimates depend from where the samples were taken. Finally, flow depth

<table>
<thead>
<tr>
<th>No.</th>
<th>Core / section name</th>
<th>Depth, m</th>
<th>Stratigraphy and lithology</th>
<th>Material</th>
<th>14C age (BP)</th>
<th>Calibrated age</th>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
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<td>1σ (68.2%) interval, BC/AD</td>
<td>Mzs, cal BP</td>
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<td>Ch-10-21</td>
<td>2.10-2.20</td>
<td>Peat covering overbank fines</td>
<td>Peat</td>
<td>7450±320</td>
<td>6660BC-5980BC</td>
<td>8330±350</td>
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<tr>
<td>2</td>
<td>Gn-10-02</td>
<td>4.00-4.30</td>
<td>Loam of channel bar swale fill</td>
<td>Macro-fossils</td>
<td>10120±70</td>
<td>10010BC-9550BC</td>
<td>11730±170</td>
</tr>
<tr>
<td>3</td>
<td>Gn-10-02</td>
<td>5.60-5.70</td>
<td>Gravely sand of channel-lag (basal) facies</td>
<td>Wood</td>
<td>9460±90</td>
<td>9120BC-8620BC</td>
<td>10770±170</td>
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<td>4</td>
<td>Gn-10-02</td>
<td>8.00</td>
<td>Alluvio-lacustrine silt</td>
<td>Wood</td>
<td>21500±650</td>
<td>24600BC-23130BC</td>
<td>25810±720</td>
</tr>
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<td>5</td>
<td>Gn-11-10</td>
<td>12.00-13.00</td>
<td>Silt of point bar swale fill</td>
<td>Wood</td>
<td>1750±250</td>
<td>1AD-560AD</td>
<td>1720±260</td>
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<td>6</td>
<td>Gn-11-02</td>
<td>7.40-7.50</td>
<td>Silt of point bar swale-fill</td>
<td>Wood</td>
<td>780±40</td>
<td>1220AD-1280AD</td>
<td>710±30</td>
</tr>
<tr>
<td>7</td>
<td>Gn-07-05</td>
<td>2.56</td>
<td>Overbank fines</td>
<td>Charcoal</td>
<td>910±80</td>
<td>1030AD-1190AD</td>
<td>830±70</td>
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<tr>
<td>8</td>
<td>Gn-07-03</td>
<td>2.63-2.70</td>
<td>Buried soil in overbank alluvial fines</td>
<td>Soil humus</td>
<td>6540±70</td>
<td>5620BC-5580BC</td>
<td>7450±70</td>
</tr>
<tr>
<td>9</td>
<td>Gn-07-01</td>
<td>5.00-5.10</td>
<td>Sand of point bar facies</td>
<td>Wood</td>
<td>2370±80</td>
<td>740BC-370BC</td>
<td>2460±140</td>
</tr>
</tbody>
</table>

Tab. 2: Radiocarbon ages from the Chekulino and Gnezdovo sites.
Tab. 2 : Âges radiocarbone des sites de Chekulino et Gnezdovo.
can be estimated from paleochannel sections at exposures or from the thickness of alluvial deposits (Maizels, 1983). Maximum bankfull channel depth $D_{bf}$ may be estimated by the total thickness of laterally accreted alluvial deposits from their top to the scoured base of channel fill (Bridge, 1978, 2003). Assessment of mean depth $D_{av}$ needs some assumptions on the $D_{bf}/D_{av}$ ratio, which may vary considerably both from section to section and from river to river. Given the above uncertainties we used the following assumptions in our palaeohydraulic estimations:

- numerical approaches were exploited with simplest possible structure and minimal set of parameters;
- ratios of paleo (index $P$) to modern (index $M$) characteristics were predicted rather than their absolute values.

The latter principle allowed omission of formulation of hypothetical relations, such as the relation between $D_{pm}$ and $D_{am}$, on the assumption that relations are valid both for the modern river and for palaeo-streams. Therefore if $D_{pm}/D_{am} = D_{bf}/D_{av}$ then:

$$D_{bf}/D_{av} = TA_{pbf}/TA_{am}$$

(eq. 1), where $TA_{am}$ is the total thickness of laterally accreted channel sediments at terraces of different ages.

The same approach was used in palea velocity assessment, for which many equations linking flow velocity and grain size have been proposed (Maizels, 1983). However, if an equation gives any systematic errors because of unknown local-specific reasons, these errors may be eliminated from ratios of equations for the same river with palaeo and modern parameters. To estimate $V_p$ we applied the Mirtskhulava (1988) equation to our non-cohesive sediments with particle size $d<0.25$ mm: $V_p = (g^2dH)^{1/4}$. Then the palaea-velocity to modern velocity ratio reads:

$$V_{p}/V_{M} = (d_p/d_M)^{1/4} (D_{bf}/D_{av})^{1/4}$$

(eq. 2),

where $d$ is sediment particle diameter taken in mm and $D_{bf}/D_{av}$ may be substituted by $TA_{pbf}/TA_{am}$ (equation 1). For homogeneous sediments defined as that for which the $d_{50}/d_{10}$ ratio is less than 5, $d$ is median diameter $d_{50}$ and for heterogeneous sediments ($d_{50}/d_{10} > 5$) median diameter of the surface pavement should be taken (Mirtskhulava, 1988). In the absence of direct measurement, Zhuravlev (1988) recommends using $d_{50}$ in the Mirtskhulava's formula in case of heterogeneous sediments. Grain size distribution for palaea velocity estimation was measured by dry sieving with an Analysette 3 PRO vibratory sieve shaker. Fractions finer than 0.05 mm were measured with an Analysette 22 laser particle size.

Following this, bankfull water discharges $Q_{bf}$ may be estimated from cross-section area defined by channel width $W$, average bankfull depth $D_{bf}$, and mean flow velocity: $Q = WD_{bf}V_{p}$. For bankfull discharges, which are often considered as channel-forming discharges, sediment entrainment flow velocities $V_{p}$ can be taken that maintain bedload transport and channel deformation. Also, bankfull discharges are usually close to mean maximum discharges. Then the ratio of palaeo and modern discharges reads as follows:

$$Q_{bf}/Q_{Mbf} = (W_{p}/W_{M}) (D_{bf}/D_{av}) (V_{p}/V_{M})$$

(eq. 3),

where $D_{bf}/D_{av}$ may be substituted by $TA_{pbf}/TA_{am}$ (equation 1).

Estimation of channel slope for palaeoflow was made on the basis of the Chezy-Manning formula: $V = n^2D^{2/3}S^{1/2}$, where $n$ is Manning's unitless roughness coefficient and $S$ is channel slope. As sediment grainsize and flow velocities did not change much (see section 4.5), both particle and bedform components of roughness must have not changed much either and therefore $n$ may be treated as a constant. Given that in case of low river sinuosity channel slopes $S$ may be substituted by valley slopes $S_v$, the ratio of palaeo and modern slopes reads:

$$S_{p}/S_{Mbf} = S_p/S_{av} = (V_p/V_M)^{2} (D_{bf}/D_{av})^{1/2}$$

(eq. 4).

3.4 - DISCRIMINATION OF RIVER CHANNEL PATTERNS

As was first shown by Leopold & Wolman (1957), meandering and braided rivers are characterized by different relations between discharge and channel slope, braiding occurring at higher discharges and slopes. In this study we exploited the latest findings by Kleinhans & van den Berg (2011) that account for channel sediments grain size and use variables independent of channel type: valley slope $S_v$ rather than channel slope, mean annual flood $Q_{bf}$ rather than bankfull discharge, reference channel width $W = \alpha Q_{bf}^{1/2}$ predicted by a hydraulic geometry relation rather than actual channel width. Channel patterns are discriminated using the median grain size of the river bed $d_{50}$ and the potential specific stream power $\omega = \rho g Q_{bf} S_{v} W_{f}^{-2/3}$, where $\rho$ is water density, $g$ is acceleration due to gravity and $\alpha$ equals 4.7 for sand ($d < 2$ mm) and 3.0 for gravel ($d > 2$ mm).

Kleinhans & van den Berg (2011) propose a succession of channel types ranked by their lateral mobility and suggest critical $\omega$ values that discriminate consecutive channel types:

- inactive (immobile) channels (indicated by subscript $i$);
- active meandering channel with scroll bars (subscript $s$);
- active meandering channel with chute bars (subscript $c$);
- braided channel (subscript $b$).

Critical $\omega$ values that discriminate consecutive channel types are proposed as follows: $\omega_{si} \approx 90d_{50}^{0.42}$, $\omega_{si} \approx 285d_{50}^{0.42}$, $\omega_{ic} \approx 90d_{50}^{0.42}$ (Kleinhans & van den Berg, 2011). More generally, the first equation is the discriminator between inactive and active channels and the last one discriminates meandering and braiding channels.

Common structure of the critical equations allows us to use the proportionality coefficient (we called it channel-type discriminator – $CTD$) to check the accordance between the estimated palaeodischarge and palaeoslope and observed palaeochannel type:
\[ CTD = \frac{\omega}{d_{50}^{0.42}} = \rho g Q_{m}^{0.5} S_{v} (\alpha d_{50}^{0.42})^{-1} \]  

(eq. 5)

\( CTD \) takes on the following critical values that discriminate different channel types:

\[ CTD_{c/s} = 90, \quad CTD_{s/c} = 285, \quad CTD_{c/b} = 900. \]

Estimation of the \( CTD \) coefficient for palaeochannels and its comparison to the above critical values provided us with possibility to verify our palaeodischarge and palaeovalley slope estimations from the point of their appropriateness for the observed palaeochannel morphology.

4 - RESULTS

4.1 - MORPHOLOGY AND SPATIAL DISTRIBUTION OF LARGE PALAOCHANNELS

In the study area, the former occurrence of flow with discharges much higher than that of the present Dnieper may be deduced from a variety of morphological evidence. First, large-scale ridge-and-swale topography has been recognized, highlighting lateral accretion of channel point bars (fig. 5A). Ridge-swale patterns may be either curved, indicating migration of large meanders, or straight, evidencing lateral shift of straight channels, or a multi-thread channels with elongated bars. In figure 5A, the average width of former scroll bars within large meanders estimated as the distance between the tops of successive ridges is about 130 m. According to the “two-thirds rule” (see section 3.1), the palaeochannel bankfull width can be estimated at ca. 200 m, which is three times that of the modern channel at the site (60-70 m). This value may indeed be underestimated, as point bar width is usually greater than the distance between scroll bar ridges.

Another type of large flow track is exhibited by wide braided palaeochannels (fig. 5B). There are usually 2-3 branches within the up to 1 km wide channel divided by wide islands. In reaches confined by high banks or valley sides or by residual hills of glacigenic material within the valley, braided palaeochannels change to single-thread channels 400-500 m wide, considerably exceeding the width of the modern river.

In many places, palaeochannels extend around high residual hills, typically 1-3 km long, with tops 20-35 m above the present river. Three groups of hills may be distinguished along the Dnieper (fig. 2). The upstream...
group of hills is preserved between the Vop’ and the Hmost’ tributary confluences, mainly on the left bank. The second group is located upstream from Smolensk at the steep bend of the valley where it changes its direction from SW to NW. The westernmost group of hills is found downstream from Smolensk. Hill dimensions, planform and orientation relative to the valley axis are highly variable. Hill planform may be rounded or elongated (teardrop-like islands) or they may have an irregular planform; some of them are normal to the valley axis, which excludes their shaping by the Dnieper.

Sites for detailed field studies were chosen in order to characterize different morphological types of large palaeochannels and traces of their lateral migration. Preserved palaeochannels that separate intra-valley hills were studied at Soloyovo (braided palaeochannel) and Chekulino (single-thread palaeochannel). The third study site is a low terrace with traces of a large migrating braided channel located near Gnездово (downstream from Smolensk).

4.2 - CHEKULINO

The Dnieper valley at Chekulino shows a NE to SW oriented residual hill 4.5 km long and up to 1.1 km wide cut into two unequal parts by the modern river (fig. 6A). The top of the larger, southern part of the hill is located at 193 m a.s.l. (33 m above the river). Two pits were excavated on the hilltop to study its sedimentological composition (fig. 6B). Four units were recognized in Pit Ch-11-04: the lower unit (unit 1, between 1.85 and 2 m depth) is composed of reddish-brown compact loam with angular gravel up to 7-8 cm clast diameter. It corresponds with the glacial till (moraine) of the Moscovian (Late Saalian, MIS 6) glaciation, according to the current generation of geological maps (Barashkova et al., 1998). Unit 2 (1.60-1.85 m depth) overlies unit 1 with clear boundary. It comprises reddish-brown (with bluish-grey spots) coarse sand with fine to angular medium gravel. This unit most likely results from fluvial-rewarking of unit 1. Unit 3 (0.95-1.60 m depth) comprises intercalations of subhorizontally laminated light-brown medium sand, bluish-grey (gleyish) silty medium sand and reddish-brown sandy loam. Sand at a depth of 1.4 m was OSL dated to 8,690 ± 570 years (GdTL-1466). The laminated and lithology of unit 3 point to sedimentation in an aqueous environment. The lamination is probably seasonal and may have been produced by redistribution of sand by rain or snowmelt waters. The upper unit 4 (<0.95 m depth) consists of well-sorted fine to medium sand, with a massive structure, reworked by pedogenesis in its upper 40 cm (unit 5). Similar massive sands 70-80 cm thick were also found above till in a pit Ch-11-01 at the edge of the hilltop. According to published sections (cores 246, 247 in Abramzon et al., 1981), this sand blanket covers not only the top of the hill, but also its sides. Its topographic position and the lack of bedding make it possible to interpret this unit as aeolian in origin, as a massive texture is quite typical for aeolian sand covers (Kasse, 2002). On the watershed south from the palaeochannel no sand blanket occurs, but a loess cover exists instead (core 249 in fig. 6C).

The surface of the palaeochannel includes two topographic steps (fig. 6C). The higher step on the left side, with a gently sloping surface at 13-14 m above the river, probably corresponds to a former side bar formed at the left bank of the river. The right bank was subject to lateral erosion, as shown by the steep slope on the southeastern side of the residual hill. The palaeochannel corresponding to the low-water stage is now a gentle hollow 300-350 m wide with its bottom at 10-12 m above the present river. The 15 m deep core Ch-11-02, located near the palaeochannel, showed 14.5 m of alluvial sands overlying dense reddish loam interpreted as the Moscovian till (unit 1). The fluvial sediments can be subdivided into two fining upward sequences each consisting of two units: ~5 m of coarse gravelly sands and 2-3 m of fine to medium sands without gravels. The lower sequence (units 2 and 3) could not be sampled because of their high water content. The top of coarser lower unit in the upper sequence (unit 4), at a depth of 3.0-3.5 m, provided an OSL age of 7,160 ± 620 years (GdTL-1465). The 2.5-m core Ch-10-21 in the talweg of the palaeochannel passed through peat and deepened into greyish loams that most probably represent overbank deposits accumulated after channel abandonment. The base of the peat, at a depth of 2.1-2.2 m, was radiocarbon dated at 7,450 ± 320 BP (GIN-14375), or 8,330 ± 350 cal yr BP. In the higher parts of the channel (core Ch-11-02) no post-abandonment fines were found probably due to the higher elevation above the river.

North of the hill, on the left bank of the modern river channel, two alluvial steps were found. The higher step, between 14.5 and 17 m above the river, corresponds to a terrace 250 m wide with its surface gently dipping towards the modern river (fig. 6C). Three metres of fluvial sediments here comprise mixed, mainly medium sands, found above Moscovian tills. OSL dating of alluvial sands in core Ch-11-05 at a depth of 1.6-2.0 m yielded an age of 9,090 ± 510 years (GdTL-1467). Below the terrace, the present-day floodplain is located at 7.5-10 m relative height, slightly dipping to the river. Core Ch-11-03 revealed the presence of 6 m fluvial sediments accumulated above the glacigenic deposits. The alluvial/glacial boundary lies at 1.5 m above the river. Large boulders washed out from tills are known to occur in the bottom of the river channel. The fluvial deposits here are rather thin compared with the 11-12 m thickness found in other parts of the valley. The fluvial sequence is composed of 0.5 m of gravels, covered by 2.5-3 m of fine to medium sands, interpreted as channel deposits. The upper 2 m consist of overbank sandy loam.

4.3 - GNEZDOVO

At this site our study focused on the traces of a large palaeochannel associated with the lower (10-11 m) terrace. On the left river bank this terrace has a relief of 2-3 m, which is produced by a regular alternation of straight SW-NE alluvial ridges 50-100 m wide divided by 70-130 m wide semi-closed hollows (fig. 7A). The elevation of successive
Fig. 6: Large palaeochannels and erosional remnants at Chekulino.

A/ Geomorphological map, B/ lithological columns and photo of pit Ch-11-04, C/ geological profile.

Geomorphology: 1/ hydrographic features: Dnieper river channel, small rivers, lakes, 2/ Holocene floodplain (5-9 m), 3/ 10-12 m Late glacial-Early Holocene terrace, 4/ 13-15-m MIS 3 terrace; 5/ alluvial valley bottom reworked by LGM glacio-fluvial processes, 6/ erosional remnants, 7/ palaeochannels, 8/ flood scour pots, edges of floodplain steps, 9/ axes of channel levees (floodplain) and elongated islands (terrace), 10/ valley shoulders and valley sides, 11/ glacio-fluvial ridges (eskers), LGM, 12/ moraine/glacio-fluvial terrain of the Moscovian (Late Saalian) glaciation, 13/ geological sections (cores, pits, exposures) and their names; 14/ settlements, 15/ profile line.

Lithology: 16/ peat, 17/ clay, 18/ loam, silt, 19/ sandy loam, loamy sand, 20/ fine to medium sand, 21/ coarse sand, 22/ gravelly sand, 23/ diamicton (stony loam), 24/ loess, 25/ buried soils, 26/ pits and cores, 27/ 14C dates (cal. yr BP), 28/ OSL dates (years).


Circled numerals at lithological columns are numbers of stratigraphic units described in the text. Elevation is in metres above the present-day river at low-water stage.

Les nombres dans les cercles des colonnes lithologiques renvoient aux unités stratigraphiques décrites dans le texte. Les altitudes sont en mètres au-dessus du fleuve actuel à son niveau de basses eaux.
ridges decreases towards the NW from 13 to 11 m above the river. Hollows and ridges were interpreted as former bars and branches, respectively, of a braided channel. This topography indicates that the terrace was constructed by a braided river with a total width several times larger than the bankfull width of the modern single-thread channel in its non-meandering reaches (80-100 m).

In section Gn-10-02 (river-bank exposure expanded by hand coring; fig. 7B), the lower unit (unit 1) is more than 5 m thick (depth between 6 and 11 m). It is composed of an alternation of fine to medium well-washed sands, fine silty sands and solid greenish-grey clay with thin horizontal bedding. This unit was interpreted as a limno-alluvial deposit. Two dates were obtained from the unit: $^{14}$C date 21,500 ± 650 BP, or 25,810 ± 720 cal. yr BP, at depth 8.0 m, and OSL date 21,400 ± 2,800 years ago at depth 9.4 m. Though the dates are not fully consistent, they lead to the conclusion that the unit was deposited during the LGM. The transition between unit 1 and the overlying sediments is sharp. Unit 2 (4.6-6 m depth) consists of an alternation of gravelly coarse sand (with medium gravels in the lower 0.25 m) and fine loamy sand, interpreted as channel deposits. The overlying unit 3 (2.5-4.6 m depth) comprises horizontal intercalations of fine sands, loamy sands and sandy loams probably deposited on the tops of channel bars. The upper 2.5 m of the section (unit 4) is overbank silty loam. Organic remains found in the channel deposits were dated by radiocarbon to 9,460 ± 90 BP (10,770 ± 170 cal. yr BP) at a depth of 5.6-5.7 m and to 10,120 ± 70 BP (11,730 ± 170 cal. yr BP) at 4.0-4.3 m.

In the borehole profile through the northern valley bottom (fig. 7C) lacustrine clays of terrace unit 1 were found only below the edge of the modern point-bar in core Gn-11-01, at a depth of 3.7 m below the modern river. This clay is underlain by well-washed coarse sand with inclusions of fine gravel, interpreted as pre-LGM fluvial deposits. Beneath the floodplain, the LGM clays were probably destroyed by lateral river erosion: at a depth of 3-7 m below the river gravelly sands were encountered, sometimes intercalated with sandy loam, interpreted as active channel sediments and OSL-dated at 15,500 ± 1,200 years ago (core Gn-11-01, depth 13.5-14.0 m). These Lateglacial channel deposits are overlain by sediments deposited by lateral migration of the river in the second half of the Holocene. The lower boundary of the Holocene sediments typically lies 3-4 m below the present-day river level, or 1-2 m above the base of the present-day channel (fig. 7C). The oldest Holocene deposits were found in core Gn-07-03. Channel sands at a depth of 11.5-11.8 m yielded an OSL age of...
6,200 ± 340 years, but humus from a buried soil formed in the overbank fines at a depth of 2.6-2.7 m was radiocarbon dated at 6,540 ± 70 BP, or 7,450 ± 70 cal. yr BP (we are inclined towards acceptance of the latter date). The Mid-Holocene is represented by channel deposits OSL dated at 4,940 ± 700 years (core Gn-11-03, depth 9.2-9.6 m). Dates from other cores indicate that most of the floodplain sedimentation occurred during the last 2,500 years.

Two terraces were cored and dated on the right bank. Alluvial sands of the 10-11 m terrace yielded OSL an age of 11,590 ± 740 years (pit Gn-12-01, depth 1.4 m), which is close to the age estimates for the fluvial deposits of similar terrace on the left bank. The 13-14 m terrace, composed of medium to coarse sand with fine gravel inclusions, was OSL dated at 54,000 ± 3,800 years (core Gn-11-13, depth 3.2-3.7 m). In both cases the fluvial deposits are underlain by similar pale pink and green-brown clays, probably of lacustrine origin, that are probably not younger than early MIS 4-3 (> 54 ka).

4.4 - SOLOVYOVO

A large braided palaeochannel is found on the left bank of Dnieper near Solovyovo and Korovinki (fig. 8A). The palaeochannel bends around a high elongated hill 3.7 km long and 1.9 km wide, with top elevation of 198 m a.s.l., or 30 m above the river. According to previously published cores (Abramzon et al., 1981) and from our investigations, the hill is composed of light reddish-brown loam loaded with gravel clasts – the Moscovian till (MIS 6, according to Barashkova et al., 1998). Both the top and the sides of the hill are covered by 2 to 2.5-m-thick fine to medium sands with coarse inclusions near their base (fig. 8B). In two pits (Sol-11-05 and Sol 11-04) we found this sand consisting of two units similar to units 3 and 4 on the hilltop at Chekulino: the lower unit with horizontal bedding (laminated sand) and the upper unit with no lamination (massive sand). In pit Sol-11-04, the lower unit in particular shows alternation between 2-10 cm beds of pure and loamy fine sands that may probably be interpreted as seasonal pairs. The lower contacts of the loamy beds are sharp, and these then grade upwards into the sand beds. OSL dating of the laminated unit in core Sol-11-05 yielded an age of 10,050 ± 670 years. The sand-till contact was not reached, but data exist from nearby cores. In cores 324 and 325 the sand is 2 and 2.5 m thick respectively. In core 323 an anomalous sand thickness of 5 m (with gravel inclusions in the lower metre) was documented, which may result from the filling of a local topographic depression and does not reflect the typical thickness of the sandy cover.

Within the palaeochannel two topographic levels occur, lying 5.5-6.5 m and 7-8 m above the river, representing palaeochannel branches and ancient chute and scroll bars (fig. 8A). In the downstream part of the palaeochannel

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**Fig. 8:** Large palaeochannel and erosion remnant at Solovyovo.  
A/ Geomorphological map, B/ profile across the palaeochannel and moraine hill.

*Fig. 8 :* Paléochenal et relief résiduel à Solovyovo. A/ Carte géomorphologique, B/ profil à travers le paléochenal et la colline morainique. Voir figure 6 pour la légende.
all branches concentrate into a single-thread channel 450-500 m wide. The 9 m deep core Sol-11-03 was located in the higher level to assess the total thickness of sedimentary fill. The lower part of the core (> 7.5 m depth) consists of light brown loam rich in angular gravels, interpreted as the Moscovian (MIS 6) till. Above, medium sands with inclusions of fine gravel were found, overlain by fine silty sands up to 2.5 m depth. These are typical channel sediments. They are overlain by a 2.5-m-thick layer of overbank silts and loams that accumulated after the palaeochannel was abandoned. As the alluvial sands at the core location lie below the groundwater table, it was not possible to sample them for OSL (neither in the core nor in a nearby pit). Sampling was possible in pit Sol-11-06, located in a drier position on a slope somewhat above lowest level and extrapolated to the studied profile (fig. 8B). Pit Sol-11-06 documented 1.4 m of post-abandonment overbank loams and half a metre of underlying fine to medium sands, identified as active-channel sediments. The total depth of the pit, excavated to the groundwater level, was 1.9 m. OSL sample taken at a depth of 1.6 m yielded an age of 9,410 ± 700 years.

The modern valley bottom consists of the floodplain with levee-hollow topography in the range 4-6 m above the river and the 8-10 m terrace gently dipping towards the river. The floodplain sediments are 14 m thick (core Sol-11/01) and are underlain by till. The alluvial base is 7.5 m below the river level and 5 m below the bed of the modern channel. The thalweg of the buried valley is located beneath the 8-10 m terrace: core Sol-11-02 passed through 23.5 m of alluvial sands and did not reach the base of the fluvial sediments. Two lithological units were documented in this core, with the boundary between them at 160 m a.s.l., or 7.5 m below the river. The lower unit is gravelly coarse sand, the upper 15-m-thick unit is fine sand. Probably the two units belong to different incision/aggradation cycles. The buried valley is rather narrow: core 318 at the terrace backside revealed glacial till at a depth of 4 m, or 6.5 m above the river (fig. 8B). The upper sand unit in core Sol-11/02 at a depth of 3.2-3.6 m (4.5 m above the river) yielded an OSL age of 12,210 ± 950 years.

4.5 - ESTIMATION OF PALAEODISCHARGE AND VALLEY GRADIENT

Estimation of past valley slopes and palaeodischarges was based on grain-size analysis from the terrace at Gnezdovo (section Gn-10/02) where the base of the Early Holocene fluvial sediments was observed at a depth of 6 m (fig. 7B). Sediments of unit 2 (depth 465-600 cm) were interpreted as active channel deposits consisting of four layers, from which four bulk samples were taken and measured for grain size: T1 – 465-525 cm, T2 – 515-545 cm, T3 – 545-575 cm, T4 – 575-600 cm. Grain-size distributions show substantial difference between units and relatively poor sorting within each of them (fig. 9A). The basal unit (sample T4) is a mixture of medium to coarse sand and fine to medium gravel with $d_{50} = 1.0$ mm. The coarsest fraction, medium gravel (10-25 mm), constitutes almost 20% of the sediment.

To characterize the modern channel sediments five samples were collected from the channel bottom along a profile across the river (fig. 7C). The modern river samples exhibit much better sorting than the terrace samples (fig. 9B). Grain-size distributions are very uniform across most of the channel width (samples R1-R4) with medium and coarse sand fractions (0.25-0.5 and 0.5-1.0 mm) corresponding to 75-95% of the sediment. Standing out is sample R5 in the deepest part of the channel adjacent to the terrace scarp: it is a mixture of sand and gravel similar to the basal alluvial layer of the terrace (sample T4), with a slightly higher $d_{50} = 1.5$ mm but lower percentage of the coarsest fraction of medium gravel – some 5%.

The critical flow velocities were characterized using the curves with largest particle size from both sets, which are samples T4 and R5. The sediments in both samples were poorly sorted and had $d_{95}/d_{5}$ ratios of 120 and 30 respectively. Therefore, following the recommendation for heterogeneous non-cohesive sediments, the $d_{50}$ parameter was used to characterize grain size in palaeovelocity calculations (see section 3.3), which constitutes 12.5 mm and 4.5 mm for samples T4 and R5, respectively (fig. 9A and B).
Channel bankfull depth was estimated from the total thickness of laterally accreted fluvial sediments as was described in section 3.3. To allow better comparison between the palaeo- and modern flows we used data from boreholes in the Chekulino and Solovyovo palaeochannels where the palaeoriver had a single-thread pattern similar to that of the present-day river channel. Excluding the post-abandonment overbank deposits, the thickness of fluvial sediments in cores Ch-11-02 and Sol-11-03 was 6.0 m (the upper alluvial unit) and 5.0 m respectively, which leads us to use an average value of 5.5 m. The modern river is represented by the Late Holocene deposits in the cross-valley geological profile at Gnezdovo (fig. 7C). Its thickness (excluding overbank fines) is 10-12 m, or 11 m in average, leading to a $TA_{\text{cp}} / TA_{\text{cm}}$ ratio of 0.5. Using the obtained values in equations 1 and 2 leads to a ratio between palaeo and modern flow velocities $V_p / V_M$ of 1.1, i.e. paleo velocities exceeded modern flow velocity only by 10%.

To estimate palaeodischarges the following average values of bankfull channel width were used: 450 m for palaeochannels (typical value for single-thread reaches both at Chekulino and Solovyovo) and 80 m for the present-day channel (average from the 80-100 m width in the Chekulino-Gnezdovo reach and the 60-70 m width at Solovyovo). Following this, the ratios between the palaeo- and modern parameters are 5.6 for channel width, 0.5 for flow depth, and 1.1 for flow velocities. Applying equation 3 gives a $Q_{\text{cp}} / Q_{\text{cm}}$ ratio of 3.05, rounded to 3.0 for further use. Maximum discharges of 830 m$^3\cdot$s$^{-1}$ at Dorogobuzh (representative for Gnezdovo and Chekulino) and 690 m$^3\cdot$s$^{-1}$ at Smolensk (representative for Solovyovo), palaeodischarges may be estimated at 2500 m$^3\cdot$s$^{-1}$ and 2100 m$^3\cdot$s$^{-1}$, respectively. This exceeds the highest recorded discharge figures (since the early 1880s) from these two gorge stations (in 1908: 1820 and 1550 m$^3\cdot$s$^{-1}$, respectively).

The change in channel gradient since the Early Holocene may be obtained from equation 4. Substitution of $V_p / V_M = 1.1$ and $D_p / D_M = 2.0$ (from the above estimation of $TA_{\text{cp}} / TA_{\text{cm}} = 0.5$) gives $S_p / S_M = 1.60$. Given that the mean sinuosity of the modern channel in the Solovyovo-Chekulino reach is 1.27 and palaeochannels were approximately straight, the ratio for valley slopes is $S_p / S_M = S_p / 1.27S_{\text{cm}}$. From the modern valley gradient in the Solovyovo-Chekulino reach of 0.084 ‰, the palaeo-gradient may be estimated at 0.106 ‰. A reconstructed long profile based on this value is shown in figure 3.

To verify our palaeodischarge and palaeoslope estimates we used Kleinhans and van den Berg’s (2011) criteria for discrimination of channel patterns to check whether the predicted palaeo parameters fit the observed morphology of palaeochannels (see section 3.4). This approach focuses on the formation of different types of alluvial bars. Hence, to calculate the channel-type discriminator parameter, $CTD$, we used $d_{\text{so}}$ for average grain-size distributions – 0.19 mm for the Early Holocene channel sediment and 0.42 mm for the present-day channel (fig. 9C). Mean maximum discharges $Q_{\text{cm}}$, for equation 5 were derived as averages of measured modern discharges from Dorogobuzh (representative of Solovyovo) and Smolensk (Gnezdovo-Chekulino) (690 and 830 m$^3\cdot$s$^{-1}$ respectively) and estimated palaeo discharges (2,100 and 2,500 m$^3\cdot$s$^{-1}$ respectively). The results of calculations are provided in table 3.

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<tr>
<th>Parameter</th>
<th>Modern values</th>
<th>Paleo values</th>
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<td>$Q_{\text{cm}}$ m$^3\cdot$s$^{-1}$</td>
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<td>2300</td>
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<td>$S_{\text{p}}$, ‰</td>
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<tr>
<td>$CTD_{\omega}$</td>
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</table>

Tab. 3: Estimation of potential specific stream power $\omega$ and channel type discrimination parameter ($CTD$) for modern and palaeochannels at the Solovyovo–Chekulino reach of the Upper Dnieper.

The equivalent calculation for the modern river, $CTD = 121$, lies within the range 90-285 that corresponds to active meandering channels with scroll bars. This is consistent with the present-day channel pattern of the Dnieper, especially in the upstream two thirds of the studied reach, while in the remainder of the reach rather long channel segments exist that exhibit no lateral migration at present, which is obvious from vegetated channel banks and the flat topography of nearby floodplain areas. Probably this is reflected by the proximity of the $CTD$ to the lower boundary of the scroll-bar channel interval. The estimated palaeo $CTD = 366$ is above the scroll bar / chute bar channel boundary ($CTD_{\omega}$ = 285) but below the chute bar / braiding boundary ($CTD_{d_{\text{so}}}$ = 900). This is consistent with active lateral migrations of palaeochannels evident from the Early Holocene terrace morphology and from the widespread occurrence of chute bar palaeochannels such as at Solovyovo. Typical braided palaeochannels such as that at Gnezdovo must have formed under much higher $CTD$ values. Probably the presence of such palaeochannels indicates underestimation of the palaeo $CTD$, which may result from underestimation of $Q_{\text{cm}}$ or $S_{\text{cp}} / S_{\text{cm}}$ or both. However, any underestimation cannot be large because in general the estimated palaeo $CTD$ corresponds well to the morphology and migration activity of the palaeostream. Therefore the reconstructed palaeo discharges and palaeo valley slope may be treated as lower estimates. At sections confined by valley sides and/or sides of residual hills, braided palaeochannels may be concentrated into a single flow (fig. 5B, 6A and 8A). These single-thread palaeochannels were excluded from verification as the Kleinhans and van den Berg (2011) theory is valid only for unconfined channels.
4.6 - ESTIMATION OF VALLEY TILT

Valley tilt may be assessed with the help of palaeochannels as reference geomorphic levels. To achieve this one needs to compare the valley slope $S_v$ estimated for palaeochannel formation and the actual valley slope measured for the same palaeochannels. To do this correlation was based on (quasi)synchronous geomorphological elements at the edges of the 100-km valley segment between Chekulino and Solovyovo (fig. 10, tab. 4).

The right-bank 8-10 m terrace at Solovyovo and the 10-12 m left-bank terrace at Gnezdovo were formed around 11.5-12 ka. The Solovyovo terrace has a smooth surface gently dipping towards the river while the Gnezdovo terrace exhibits levee-hollow topography with relief of 1-2 m. Calculations based on average elevations give a slope of 0.060 %. For the Solovyovo and Chekulino palaeochannels, formed around 8.5-9.5 ka, two variants were assessed: correlation by the top of the alluvial sediments and correlation by the alluvial base. At Solovyovo, both characteristics were taken from core Sol-11-03 (fig. 8B). The alluvial top was measured at the top of the sands with the 2.5-m-thick overbank fines removed because they were accumulated after palaeochannel abandonment. At Chekulino the top of the channel deposits was represented by the base of post-abandonment loams in core Ch-10-21. The alluvial base was determined in core Ch-11-02, which was drilled to the top of the underlying till. The fluvial deposits-till contact could not be used for our purposes as the total thickness of fluvial sediments that fill the Chekulino palaeochannel (14 m) obviously exceeds the active sediment thickness for this palaeochannel. This assumption is based not only on the thickness value itself, but also on the fact that the fluvial bed lies 7 m lower at Chekulino than in the nearby Gnezdovo terrace in section Gn-10-02 (cf. fig. 6C and 7C). In contrast, core Ch-11-02 provided evidence for basal fluvial sediments at a depth of 8 m that overlie a 2-m-thick layer of fine well-sorted sands. This level is more likely to be correlated with the terrace bed at Gnezdovo and was accepted as the active alluvial bed.

Slope calculations using the top and bottom of the fluvial sediments in the palaeochannels yielded values of 0.061 % and 0.055 %, respectively, which are close to each other and also close to the slope calculated from the inclination of terraces (fig. 10, tab. 4). The average of the three slope estimates is 0.059 %. This value is considered as an approximation of the present-
day gradient of the Early Holocene valley bottom. It is lower than the modern valley gradient of 0.084 % in the Solovyovo-Chekulino reach, and than the valley slope $S_{p}$ that maintained formation of large palaeochannels, which was estimated at 0.106 % (see section 4.5). This indicates the occurrence of valley tilting, with relative subsidence in its upstream part. Quantitative estimation is possible for the magnitude of this subsidence. Values of valley slope and distance between Solovyovo and Chekulino indicate a fall in valley elevation of 0.106 % × 94 km = 10.0 m for the period of palaeochannel formation and 0.059 % × 94 km = 5.5 m for the present-day position of the same palaeochannels. Consequently, since the period of palaeochannel formation, the Solovyovo site has subsided by 10.0 - 5.5 = 4.5 m relative to the site at Chekulino. Given the actual fall in river elevation between Solovyovo and Chekulino, which is 7.9 = 8.0 m, one can estimate that 8.0 - 5.5 = 2.5 m of this subsidence must have been compensated by modification by the river of its long profile: by incision at Chekulino, or aggradation at Solovyovo, or both. Discussion on the above estimates will be provided in section 5.3.

5 - DISCUSSION

5.1 - AGE AND ORIGIN OF THE UPPER DNIEPER VALLEY

Kvasov (1975, 1979) proposed that the Upper Dnieper valley did not exist before the LGM and related its origin to the formation and further outflow from large proglacial lakes. He postulated the existence of two lakes: the Orsha Lake, downstream from Smolensk, with a water level at 180 m a.s.l., and the Dorogobuzh Lake, upstream from Smolensk, with a water level at 200-215 m a.s.l. It was assumed that the water levels of both lakes were controlled by the elevation of local outlets. The Dorogobuzh Lake was assumed to have overflowed into the Volga Basin and thus contributed to the Khvalynian transgression of the Caspian Sea. A large fall in water level between the two lakes promoted the formation of a spillway valley in the Smolensk reach and the ultimate drainage of the lakes, allowing the westerly-aligned reach of the valley between Dorogobuzh and Smolensk to be formed.

Our results do not confirm the above scenario. Dating of the 13-15 m terrace at Gnezdovo to MIS 3 (fig. 7C, core Gn-11-13) demonstrates that the Smolensk reach of the valley had existed long before the MIS 2 advance of the SIS. We obtained evidence for only one LGM lake (fig. 7B and 7C, core Gn-10-02) that occupied a relatively short section of the Smolensk valley reach, right at Kvasov’s divide between his two hypothesized proglacial lakes. The lake most probably formed due to glacial damming of the valley in the Chekulino reach (fig. 6A). Damming is evidenced by glacio-fluvial topography over the former valley bottom on the right-bank side of the present-day river, in particular high eskers-type sand and gravel ridges that could have accumulated only within or in the close vicinity of ice masses. Lake level was not higher than 173-175 m a.s.l. because the right-bank 13-15 m terrace at Gnezdovo does not bear lacustrine deposits above the MIS 3 alluvial sands (fig. 7C, core Gn-11-13). The lake extended upstream for no more than few tens of kilometres, which follows from the absence of the LGM lacustrine deposits in the Solovyovo reach of the valley.

To prove the existence of the Dorogobuzh Lake, Kvasov (1975, 1979) mentions the occurrence of a 30-35 m terrace upstream from Smolensk covered with 2-4 m of lacustrine sands, but he does not provide any particular sedimentological section and geomorphological profiles to support this. In our study we found sand-covered terrace-like surfaces at the tops of residual hills composed of Moscovian till at 30-35 m within the valley at Chekulino (fig. 6C) and Solovyovo (fig. 8B). These sands were interpreted as aeolian and aeolian- reworked by slopewash and were dated to the Early Holocene (see sections 4.2 and 4.4), which excludes their association with LGM proglacial lakes. On the other hand, these dates are close to alluvial dates from adjacent large palaeochannels. Following this, the most likely source of sand could be found on wide bars of adjacent large channels that were dry during low water seasons; active wind transport and the formation of aeolian dunes and sand covers near to river channels is widely known from river valleys overloaded with sand (Allen, 1965a; Kasse, 2002).

We conclude from the above that the Upper Dnieper valley was formed long before MIS 2, probably at the end of MIS 6 (Moscovian deglaciation). During the LGM, the valley was not totally occupied by large proglacial lakes, except for the relatively small glacial-dammed lake in the vicinity of Smolensk.

5.2 - INCISION/AGRADATION HISTORY OF THE UPPER DNIEPER SINCE MIS 3

Data from Gnezdovo (section 4.3) allow a chronology for valley filling and deepening during the last 50 ka to be reconstructed, by using alluvial base elevation and the top of laterally accreted fluvial sediments (excluding overbank deposits). The former indicates the position of river bottom in the deepest pools, and the latter gives the elevation of channel bars. Both parameters are plotted on the elevation versus time graph (fig. 11A) with elevation measured relative to the present-day river. Data for the plot were taken form the cross-valley geological section (fig. 7C). In the middle of MIS 3 the river was some 10 m higher compared to the present, which is exhibited by the 13-15 m right-bank terrace dated around 54 ka. Evidence for the pre-LGM deep incision is provided by the position of the bed of LGM lacustrine clays, which accumulated in the bottom of the channel. The LGM was marked by a > 10 m valley filling by fluvio-lacustrine deposits. The presence of fluvial sediments interbedded within the lacustrine deposits allows this filling to be located in the vicinity of the channel bottom.

The LGM fluvio-lacustrine deposits in section Gn-10-02 are overlain by the Early Holocene alluvial sequence of the 10-12 m terrace related to formation of large palaeochannels. However, before the formation of this terrace
one more incision-aggradation cycle had occurred. The presence of coarse channel sands dated to 15.5 ka in core Gn-11-01 actually demonstrates that before this time the river was incised deeper than the present-day channel. Then aggradation took place and by 12-11.5 ka the 10-12 m terrace was deposited with the alluvial base lying 4 m above the modern river. This incision-aggradation event detected at Gnezdovo is consistent with the excessive alluvial thickness of the 8-10 m terrace at Solovyovo (core Sol-11-02) which provides evidence for aggradation prior to the formation of the large palaeochannel. It also explains the large thickness and two-unit composition of the fluvial sediments that fill the Chekulino palaeochannel.

The final incision down to the modern position is dated at Gnezdovo between 11.5 ka (formation of the left-bank terrace) and 7.5 ka (buried soil in overbank fines in core Gn-07-03). Given the time needed for accumulation of alluvial section and soil formation (core Gn-07-03), and taking into account the chronology of abandonment of palaeochannels at Chekulino and Solovyovo that were active prior to this incision, the incision may be bracketed between 9 and 8 ka. The magnitude of the incision differs considerably when considering the top or base of the fluvial deposits, with the range from 1-2 m to 8-9 m respectively. This is due to channel pattern transformation associated with the incision: the Early Holocene braided river transformed into a single-thread channel and become narrower but much deeper. Since that time the river long profile may be thought to have reached the quasi stable state. Incision/aggradation magnitudes did not exceed 1-2 m.

5.3 - PROGLACIAL CONTROL ON VALLEY DEVELOPMENT

The glacial forebulge uplift that accompanied the advance of the SIS in the first half of MIS 2 is likely to have changed the Upper Dnieper valley slope. Predicted LGM palaeotopography (fig. 4) gives an estimation of the LGM valley gradient in the westerly-aligned reach between Dorogobuzh and Orsha twice the present-day value (tab. 5). Increasing downstream gradient must have resulted in raised flow velocities, potentially contributing to the pre-LGM deepening of the valley (fig. 11A). In contrast, the southerly-aligned valley reach downstream from Orsha was on the uplifting proximal side of the forebulge and must have experienced a decrease in valley gradient. According to the output from the ICE-5G model (fig. 4), the relative rise at the downstream point, the city of Rogachev, would have exceeded the fall in valley elevation between Orsha and Rogachev.
and produced a reverse gradient and thus an impediment river flow (tab. 5). This kind of river damming is likely to explain the wide occurrence of MIS 2 lacustrine deposits in the Dnieper, Berezina and northern Pripyat’ basins in central Belarus (Kvasov, 1975, 1979; Kallieki, San’ko, 1992, 1998; Karabanov & Matveyev, 2011). Drainage in these basins was generally directed southwards and must have experienced backwater effect due to forebulge uplift (fig. 4). Forebulge subsidence provided drainage amelioration, which is exhibited in increase of valley gradient between Orsha and Rogachev (fig. 11B).

In the Dorogobuzh-Orsha reach, forebulge collapse must have been followed by a decrease of valley gradient and resultant channel aggradation. Following the ICE-5G model (Peltier, 2004), valley back-tilting was marked in the interval 21-9 ka and decelerated thereafter (fig. 11B). However our results would fit better with delayed forebulge subsidence in comparison with that predicted by ICE-5G. First, the deep post-LGM incision at around 15.5 ka requires persistence of a steep valley gradient at that time. Second, we have estimated valley tilting by ca. 4.5 m in the Solovyovo-Chekulino reach during the last 9-8 ka (after abandonment of large palaeochannels; see section 4.5). In the Dorogobuzh-Orsha reach, which is 2.5 times the length of the Solovyovo-Chekulino reach, the amount of tilt could be around 10 m, which is noticeably higher than the tilt of ca. 3 m predicted for the last 9 ka by ICE-5G (fig. 11B). Also, the occurrence of rapid pre-Holocene aggradation is more readily understood if it was coincident with a rapid decrease in valley gradient, which is not the case according to ICE-5G (cf. fig. 11A and B).

The last incision phase, between 9 and 8 ka, which coincided with the abandonment of large palaeochannels, is difficult to explain in terms of glacio-isostatic forcing. Forebulge subsidence still continued after 9 ka, and the river should have aggraded in response to decreasing valley gradient. The possible reason for incision could be the upstream propagation of regressive erosion from the Orsha-Rogachev reach, where forebulge collapse must have resulted in consistent increase of valley gradients (fig. 11B) accompanied by stream incision. However the convexity of the river long profile at Orsha (fig. 3) shows that this incision wave has not yet arrived in the Smolensk reach of the river. Therefore the reason for the incision may lie in climatic forcing or some intrinsic processes, which is discussed below (section 5.4).

Discussion on the role of glacio-isostatic control should account for possible contribution from other types of tectonic movements. Crustal uplift and subsidence reversals were reported from the Don and Dnieper river basins based on analysis of river terraces (Matoshko et al., 2004; Westaway & Bridgland, 2014). However the timescales of these reversals, hundreds of thousands – millions of years, are far above the timescale of valley processes reported in the present study. No active faults were documented in the region by geological surveys (Abramzon et al., 1981; Barashkova et al., 1988), therefore differential tectonic movements may not be regarded as the cause for the envisaged valley gradient changes.

5.4 - CLIMATIC FORCING AND PALAEOHYDRO-LOGICAL CHANGE

The LGM climate at 21 ka was cold and dry over the whole of Northern Eurasia (Kageyama et al., 2001). This promoted river aggradation, which is detected almost everywhere in Central and East Europe (Starkel et al., in press). In the Upper Dnieper such climatic conditions coincided with glacial damming of the river, which probably contributed to the fluvial aggradation observed at Gnezdovo (see core Gn-10-02 in fig. 7B,C; fig. 11A). Klimanov (1997) found evidence for a moderate precipitation decrease in the Upper Dnieper region in the Allerød and a considerable decline in precipitation in the Young Dryas. In the Upper Volga region the change from dry and cold to a warmer and more humid climate occurred around 15 ka, but the interval between 13 and 11.5 ka was markedly dry (Klimanov, 1997; Khotinsky & Klimanov, 1997), leading to lowering of lake levels (Wohlfarth et al., 2007). Climatic reconstructions for North-Western Russia have pointed to increasing humidity since the beginning of the Holocene, but in the interval 10.0-8.0 ka low fluvial discharge has been predicted (Wohlfarth et al., 2007). Khotinsky & Klimanov (1997) also report Early Holocene precipitation below
present-day values, but suggest constant increase between the start of the Holocene and 8.5 ka, when precipitation was comparable with that of the present day (Velichko et al., 1997). It is likely that a combination of lower temperatures and such precipitation values would lead to higher discharge than at present. This is supported by the findings of Starkel (1999), who recognized the global occurrence of a cooler and wetter climatic phase between 9.5 and 9.0 ka. This humid phase corresponds with the final stage of large palaeochannel formation in the Upper Dnieper, while their initiation during 12-10 ka correlates with predicted lower humidity conditions.

Almost no palaeoclimatic signals have been reported that could have caused fluvial during 9-8 ka, the exception being the moderate peak in precipitation at 9-8.5 ka. Notwithstanding the fact that the cited palaeoclimatic estimates were based on statistical processing of pollen data, which may not be sufficiently sensitive to assess precipitation variations in humid regions (Guiot et al., 1993), the changes are in any case too small to have influenced the fluvial response detected in the Upper Dnieper.

The other approach to estimation of discharge changes is the study of geomorphic traces of streams and their evaluation in terms of palaeodischarge. Geomorphic analysis reveals the wide occurrence in European river valleys of large meandering palaeochannels that were created by extremely high discharge events. In the central part of the East European Plain large meanders developed between 18 and 13 ka (Borisova et al., 2006; Sidorchuk et al., 2009). A post-LGM increase in discharge is supported by frequency analysis of absolute chronology data classified according to the magnitude of fluvial activity (Panin & Matlakhova, 2014). Abandonment of large meanders in Central Europe was dated in the range 16-12 ka: in Poland and Hungary they developed following the Bølling and up to the end of the Younger Dryas (Szułanski, 1983, 1986; Vandenberghhe et al., 1994; Gebcica et al., 2009; Kasse et al., 2010), in Lower Danube tributary valleys they have been dated at about 15.5 ka (Howard et al., 2004). The age of large meanders and associated high discharge episodes corresponds well with incision by the Upper Dnieper, also dated at around 15.5 ka, which might mean that this incision was not only a response to a glacio-isostatically driven increase in valley gradient (see section 5.3) but also a response to higher fluvial discharge.

In the Early Holocene, there were extremely variable but generally high levels of fluvial activity in the central East European Plain until ca. 8.5 ka (Panin & Matlakhova, 2014). This is consistent with the formation of large palaeochannels in the Upper Dnieper, which occurred during approximately the same time period. However, the Dnieper palaeochannels have no analogues in the East European Plain in terms of size and estimated palaeodischarge (see section 4.5). The fluvial incision and the transformation from braided to single-thread channel pattern that occurred between 9 and 8 ka do not correlate with any known high-magnitude palaeoclimatic or palaeohydrological events. On the contrary, they correspond with declining fluvial activity at around 8.5 ka and to the beginning of the Holocene longest low-flood period between 8.5-5.5 ka (Panin & Matlakhova, 2014). Given that the glacio-isostatic control acted against than promoting river incision (see section 5.3), valley deepening during 9-8 ka may be explained by intrinsic mechanisms – namely the transformation of shallow braided streams into deep single-thread channels.

5.5 - ORIGIN OF LARGE PALAEOCHANNELS AND RESIDUAL HILLS

5.5.1 - Chronology of large palaeochannels

The chronology of the large palaeochannels of the Dnieper River has been improved by numerical dating. In both the Chekulino and Solovyovo palaeochannels, sands with gravel inclusions were found that could not have originated from post-abandonment aeolian or alluvial overbank deposition. In the Upper Dnieper, overbank sediments consist of loams, silts and fine sands with no gravels, which is why the thick (several metres) sandy deposits that have been found in palaeochannels, containing layers of coarse sand (1-2 mm) and inclusions of gravel (1-3 cm), are interpreted as active channel alluvial facies. Reworking of fluvial sediments by the active channel of the Dnieper after a decrease in discharge, or by small tributaries using the palaeochannels after their abandonment by the main river, can also be excluded since coring locations were chosen at places where no traces of small channel migration were found. Slope mass movements can also be excluded on account of the facies and the location of cores, which was well away from valley- or hill-sides. This demonstrates that the ages obtained correspond to active channel sedimentation that took place shortly before abandonment. Additional indirect support of Early Holocene activity in large palaeochannels is provided by the Early Holocene dates from hilltop sands interpreted as water-reworked aeolian sediments derived from vast channel bars (see section 5.1).

Additional data related to palaeochannel chronology is provided by the 10 m terrace at Gnedzovo (section 4.3; fig. 7B). Large levee-hollow systems provide evidence for lateral migration of a large braided palaeochannel, which allows the formation of the terrace to be related to the palaeochannel activity at Chekulino and Solovyovo. The channel deposits, including the lag facies, were dated to the Early Holocene, both in section Gn-10-02 by radiocarbon and in section Gn-12-01 by OSL. Following from this, the large Dnieper palaeochannels cannot be associated with glacial meltwater; moreover, they are younger than those found in other European regions and dated to the Lateglacial.

This result is consistent with those obtained in the Belarusian reach of the Dnieper valley between the Orsha and Rogachev; in that area, large palaeochannels have been observed and correlated with the 10 m terrace, which was radiocarbon dated in the valley of the Chizhovka, a Dnieper tributary near Orsha (Kalicki & San’ko, 1992, 1998). A radiocarbon date of 17,150 ± 300 BP in the Chizhovka section was obtained from lacustrine silts that underlie the
3-m-thick alluvial gravels of the terrace. It follows that the age of the alluvial gravels and of the terrace is younger. Our results are also consistent with those of Matoshko (2004), who described large braided palaeochannels in the middle Dnieper and Prypiat’ valleys, located between a terrace and the floodplain. As the terrace is dated to the Pleistocene/Holocene boundary (see table 1 in Matoshko, 2004), the high discharges that produced large meanders may be attributed to the beginning of the Holocene, which is in agreement with the Early Holocene increased discharge recognized here for the Upper Dnieper.

5.5.2 - Braiding pattern of large palaeochannels

The formation of large channels needs specific hydraulic conditions that are different from those of the present day. It was estimated in section 4.5 that the large Dnieper palaeochannels were formed with a three-fold increase of flood discharge and a 30% increase of valley gradient. Although it has been successively modeled in laboratory flumes (Schumm & Khan, 1972), the formation of a braided channel due to an increase of downstream valley gradient is rather rare in natural rivers because significant gradient change as a result of intrinsic factors such as valley aggradation requires a significant period of time. However it is possible under extrinsic control, such as when a river valley is tilted due to crustal warping. Therefore, the question arises as to why the Early Holocene Dnieper was forming braided channels rather than deepening its channel, when gradient and discharge were much higher than at present.

The answer probably lies in accounting for the change of channel parameters over time rather than their absolute values. In laboratory models, sediment discharge regulation is used to prevent incision in response to an increase of valley gradient (Schumm & Khan, 1972). In the Upper Dnieper, the formation of large braided palaeochannels was preceded by significant aggradation (fig. 11A) in response to the sudden decline of downstream gradient produced by collapse of the glacial forebulge (section 5.3, fig. 11B). Declining gradient and flow velocity must have resulted in the stream becoming overloaded with sediment, leading to aggradation and widening of the channel and finally the formation of a braided pattern. Early Holocene braiding seems therefore to have been the response of the river to overloading by sediments due to a rapid extrinsically forced decline of valley gradient.

Slight incision by the Upper Dnieper may have started at the Pleistocene-Holocene boundary, as suggested by a staircase of ancient channel bars on the lower terrace at Gnedzovo: older ridges located in the southeast are 1-2 m higher than younger ridges located closer to the modern channel (fig. 7A). However this gradual deepening was not accompanied by any change of river style. The existence of an incising braided channel requires a combination of sediment-supply deficit with a valley gradient and discharge combination unfavourable to meandering. Valley gradient is a function of long-term development of the river’s long profile (incision or aggradation) and/or external forcing such as tectonic tilting. Sediment delivery is subject to short-term variations determined by climate forcing (Vandenberghe, 2003, 2008). Therefore a combination of high valley gradient with bedload deficiency is quite possible, since their time scales and governing factors are unconnected. Vandenberghe (2008) recognized limited incision by lowland braided rivers at the beginning of cold stages within glacial-interglacial cycles.

Fundamental transformation of large braided palaeochannels into single-thread small channels of the modern type corresponded to the incision event at the Early-Mid Holocene boundary at 9-8 ka (fig. 11A). The origin of this incision cannot be attributed either to crustal movements (see section 5.3) or to climatic forcing (see section 5.4). Most probably this event occurred in response to some intrinsic control. We propose that this incision was pre-determined by the exhaustion of sediment storage accumulated in the valley bottom during the previous cold epoch. Flow concentration in a single-thread channel due to reduced sediment supply could thus have been the forcing for strong deepening of the channel bottom. Also, the incision was most probably completed before the beginning of the Mid-Holocene low-flood epoch, because it coincided with such significant events as the abandonment of the Chekulino palaeochannel and the formation of the modern spillway reach of the valley, which is discussed in the next section.

5.5.3 - Origin of residual hills

The intra-valley residual hills at Chekulino and Solovyovo (fig. 6 and 8) are separated from the valley sides by Early Holocene palaeochannels, which provides a hypothesis of their interrelated origin. However the morphology of these abandoned valleys does not allow consideration of them as spillways: both the valley sides and the sides of the residual hills are usually gentle and pass smoothly into the palaeochannels. No terraces were found within the abandoned valleys except for former palaeochannel bars. Boreholes in the studied palaeochannels at Solovyovo and Chekulino were drilled through the fluvial sediments to the underlying till. All this leads to the proposal that the erosional dissection that separated the intra-valley residual ‘moraine hills’ had occurred before the formation of the large palaeochannels. Residual hills had probably been formed due to erosional dissection of the Moscovian (MIS 6) glacial plain with the formation of the Upper Dnieper valley. The large Early Holocene palaeochannels then made use of the troughs that had already existed at the sides of the valley, modifying them in just a few places by lateral erosion of the hillsides (e.g., the right bank of the Chekulino palaeochannel in the cross-profile in fig. 6C).

6 - CONCLUSION

This study has produced the following major findings: – intra-valley hills in the Upper Dnieper are a part of a complex ancient landscape produced by erosional dissec-
tion and formation of residual “moraine hills” elongated along the valley. This probably took place during the initial formation of the valley at the end of MIS 6. This landscape has been progressively buried by alluvial sediments through the Lateglacial and the Holocene. Hollows separating residual hills were subject to overbank fluvial deposition before the onset of the Holocene and some of them were reoccupied by the Dnieper in the Early Holocene;

- large palaeochannels were active in the Early Holocene between 12 and 8.5 ka. They were formed by flood discharges ca. three times larger, and with a valley gradient some 30 % higher, than those of the present-day river. It follows from their age that they cannot be related to the outflow of LGM glacial meltwater;

- valley evolution during and following MIS 2 was strong influenced by proglacial effects – glacial damming and valley tilting due to forebulge development. Increased valley gradient caused deep pre-LGM incision, which was interrupted by glacial damming 15 km downstream from Smolensk. This caused rapid aggradation around the LGM, with incision back to the previous level in the Lateglacial. Significant aggradation occurred between 15.5 and 12 ka as a fluvial response to valley back-tilting caused by forebulge collapse. Valley back-tilting continued in the Holocene with a total amplitude of ca. 4.5 m along a 100-km valley reach. This is larger than predicted by the ICE-5G (VM2) hydro-glacioisostatic model (Peltier, 2004) and may probably be considered as delayed subsidence of the forebulge in comparison with the modelling prediction;

- palaeohydrological changes deduced from the regional data acted in conjunction with the glacio-isostatic forcing. The three-fold rise of discharge estimated for the large palaeochannels indicates a large increase in runoff over the Upper Dnieper catchment in the Early Holocene;

- deep river incision to the modern levels occurred between 9 and 8.5 ka. It can be explained by intrinsic rather than external forcing; possibly a decrease of stored alluvial sediment limited bedload supply and stimulated the concentration of the river in a narrow but deep single-thread channel. However the incision could have been triggered by extreme flooding events.

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