Chapter 14
Floodplain Connectivity

József Dezső, Dénes Lóczy, Ali Mohamed Salem and Gábor Nagy

Abstract Floodplains fulfil vital ecosystem services (supporting water management, biodiversity, agricultural production, ecotourism and others). Since a satisfactory water supply is indispensable for the provision of such services, in addition to longitudinal channel connectivity, lateral channel/floodplain hydrological connectivity is of primary importance. As a consequence of river regulation, however, floodplains shrunk considerably in area, 'protected' floodplains connection with the river channel which had produced them and became severely threatened ecosystems. In the Drava Plain, too, disconnected (or 'geographically isolated') oxbows became typical. With reduced surface connectivity, groundwater flow becomes the main driver of connecting processes (profundal type of oxbow). Effective porosity and hydraulic conductivity of alluvial deposits and seepage from an oxbow lake (the degree of clogging of floor deposits) were calculated to estimate groundwater movements and to reveal water exchange between oxbow lakes and the active river channel. Subsurface connectivity under drought conditions was simulated by hydrological modelling with the help of HYDRUS-1D and MODFLOW 6 packages. Planning rehabilitation efforts subsurface connectivity too should be considered.

Keywords Channel/floodplain connectivity · Land drainage · Groundwater flow
Hydraulic conductivity · Seepage · Hydraulic modelling · Rehabilitation

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14.1 Introduction

In geomorphology and landscape ecology, the concept of connectivity is applied in three different meanings (Croke et al. 2013):

(1) *landscape (topographic) connectivity* denotes the physical coupling of landforms (e.g. hillslope to channel) within a drainage basin (e.g. Michaelides and Wainwright 2002; Brierley et al. 2006);
(2) *hydrological connectivity* refers to the communication of landforms manifested in surface and subsurface water flow (e.g. Ambroise 2004; Bracken and Croke 2007; Opperman et al. 2010) and
(3) *sedimentological connectivity* relates to the movement of sediments and attached pollutants between landscape units (e.g. Hooke 2003; Fryirs et al. 2007; Fryirs 2013).

Although the term is widely used, confusion remains on how to quantify connectivity at various spatial and temporal scales (Croke et al. 2013).

Even before the concept of hydrological connectivity became accepted, it had been pointed out that, nutrients, sediments and organic matter carried by water move laterally and are deposited on floodplain surfaces (Gregory et al. 1991; Zwoliński 1992). A central theme in large river ecosystem functioning has been the flood pulse concept (Junk et al. 1989; Sparks et al. 1990; Tockner et al. 2000), which holds that, during floods of proper level and duration, floodplain connectivity allows the exchange of nutrients and feeding and spawning for some fish species in the floodplain. In the often heavily altered flow regimes of regulated rivers, however, flood pulses, which had formerly ensured connectivity, are reduced in magnitude, frequency and duration (APFM-WMO 2017).

Landscape ecological studies underline that longitudinal hydrological connectivity is a key property for the survival of the whole system (Tockner and Stanford 2002; Brierley et al. 2006). The term does not only refer to the river channel, but to the riverine corridor, too, and it denotes a central concept of restoration/rehabilitation projects (Piégay et al. 2000; Sulc-Michalková and Sulc 2011).

In addition to surface connectivity, water may also connect with the floodplain and its wetlands through groundwater flow (Brinson et al. 1995; Jacobson et al. 2011). Unconsolidated and permeable floodplain deposits facilitate rapid and dynamic connections between river channels and wetlands adjacent to the channel (Amoros and Bornette 2002) or even at considerable distance from that (in 'geographically isolated' position—Ameli and Creed 2017). (It is also observed in the case of the Drava River—Dezső et al. 2017). The hyporheic corridor may extend some kilometres away from the river channel into the alluvium (Stanford and Ward 1993) and can be an important source of water supply to oxbow lakes (Tockner et al. 1999). Other sources of groundwater are neighbouring hills and uplands, where run-off maintains hillslope/floodplain links (Kelly 2001) and establish a connectivity chain.

In view of complex interactions among surface and groundwater, topography and alluvial deposits present a challenge to the assessment of rehabilitation potential

(Fryirs and Brierley 2000, 2016). For successful interventions, detailed historical geomorphic analyses at floodplain and catchment scales are needed (Brierley and Fryirs 2005, 2008; Kondolf et al. 2007; Hohensinner et al. 2011). Analysing a Danube restoration project, Tockner et al. (1998) claim that primarily fluvial dynamics, the associated connectivity gradients and a natural disturbance regime have to be reestablished. Then the ecosystem can hopefully be maintained with only minimum effort.

14.2 Types of Hydrological Connectivity

Floodplain hydrology, morphology and hydraulic connectivity are influenced by a wide range of factors (Shankman 1993; Amoros and Bornette 2002; Hudson 2010; Kupfer et al. 2010). The stage of infilling for abandoned channels (including oxbows) depends on the time elapsed since cut-off, avulsion or bifurcation and the rate of plugging processes. The density of vegetation controls the accumulation of organic fill. The links established with cross-floodplain tributaries help abandoned channels remain hydrologically active longer (Phillips 2009). Additional local processes can also affect the evolution of floodplain depressions.

Phillips (2013) identified six different modes of surface hydrological connectivity for floodplain wetlands (oxbow lakes, sloughs and paleochannels):

(1) *flow through* (regular river flow to and from the main channel);
(2) *floodchannel* (flow at high water stages, partly reaches the main channel);
(3) *fill and spill* (flow at high stages, fill to a threshold level and then overflow into flood basins);
(4) *fill and drain* (fills at high river discharge and returns to the main channel);
(5) *tributary occupation* of former river reaches and
(6) *disconnected* (no flow except during large floods).

He found that lateral distance from the active channel is poorly related to hydrological connectivity.

In their lake typology, Dawidek and Ferencz (2014) also claim that 'the hydrological (the degree of filling of the basin) and ecological state mostly depends on the type of connections that the lake has to the parent river'. Dawidek and Turczyński (2006) identify four types of connection between floodplain lakes and the main river (Fig. 14.1):

(1) *confluent* (the floodplain lake receives water from the river in the direction of general slope of the valley);
(2) *contrafluent–confluent* (the lake is fed by the river from both upstream and downstream directions);
(3) *contrafluent* (water supply during floods and backflow to lake at the same site) and
(4) *profundal–confluent* (the lake is fed primarily from groundwater and secondarily from the river channel).

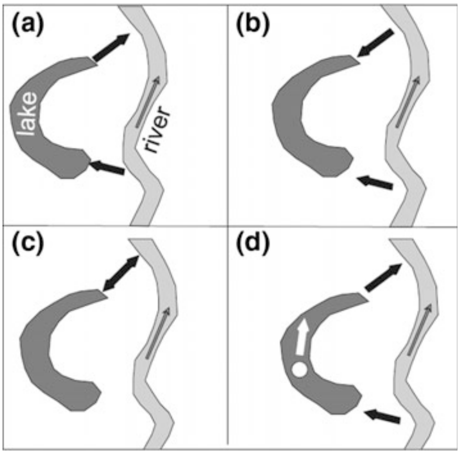

Fig. 14.1 Types of oxbow lake recharge from the main river channel (on the example of the Bug River, Poland—after Dawidek and Turczyński 2006). A, confluent; B, contrafluent–confluent; C, contrafluent; D, profundal–confluent

Naturally, there is an extreme profundal type, too, where groundwater flow is the sole supplier of water to the lake. Even for lakes where no profundal connection is predominant, groundwater flow is important (Dawidek and Ferencz 2014). As a matter of course, the rate of recharge is also dependent on these types. The interception and transpiration of riparian forests can significantly modify the water balance of floodplains (see Chap. 12 and Piégay et al. 1997).

The EU Water Framework Directive (WFD) (European Commission 2000) also acknowledges that surface/subsurface water interactions play a crucial role in the water budget of floodplains for the restoration of floodplain habitats (Downs and Thorne 2000). Major streams, like the Drava River, and their hyporheic zones maintain a hydraulic balance with groundwater. Among groundwater bodies and aquifers, the unconfined aquifer reacts most rapidly to rainfall events. Therefore, groundwater and surface water have to be conceived as components of a single system (Winter et al. 1999). Although the protected floodplain is only exceptionally covered by water along regulated rivers like the Drava, inundations in the perirheic zone (as defined by Mertes 1997) are regularly observed during wet spells.

Along the Drava River, flood-control structures also disrupted links on the surface and reduced connectivity to groundwater flow. Parafluvial and floodplain hyporheic flows also seem to be inhibited. Increasingly, severe drought conditions are indicated by dropping water levels recorded in groundwater observation wells even at 2–3 km distance from the channel.

14.3 Methods and Discussion

There are analytical and synthetical approaches to quantify floodplain/channel connectivity (also embracing flood risk). Studying surface connectivity, Croke et al. (2013) identified macrochannel width as a crucial factor. Macrochannels and associated landforms (within-channel benches, macrochannel banks and floodplains) were represented on fine-resolution digital elevation models based on LiDAR survey. Bankfull Average Recurrence Interval (ARI) was computed for channel reaches of expansion and contraction using the one-dimensional (1D) flow hydraulic model HEC-RAS. Valley floor width was established from the valley bottom flatness index (MrVBF—Gallant and Dowling 2003). Significant nonlinear changes in channel capacity were found to control the spatial pattern of hydrological connectivity. A network index (NI) has also been proposed to predict surface connectivity in agricultural catchments (Lane et al. 2004; Shore et al. 2013).

An index for river/floodplain connectivity, called the Land Capability Potential Index (LCPI), was suggested to assist regional-scale restoration planning of agricultural land along the Lower Missouri River (Jacobson et al. 2007, 2011). The LCPI integrates modelled water-surface elevations, floodplain topography and soils to index relative wetness of floodplain patches. Schwarz et al. (1996) assessed hydrological connectivity in terms of distance, elevation difference and channelized connections to the active channel. These approaches are directed at studying the fertility of agricultural land and not really appropriate for the investigation of the Drava floodplain.

To find suitable indicators for *subsurface connectivity* is even more problematic (Golden et al. 2014; McLaughlin et al. 2014). Most authors suggest the application of coupled surface–subsurface flow models, like MODFLOW (Brunner et al. 2010). The MODFLOW wetlands package and its recent modified versions (MODFLOW 6) are useful tools to depict subsurface hydrologic connectivity of wetlands, where groundwater is the dominant flow pathway. MIKE-SHE covers six key processes at the watershed scale: overland flow, channel flow, unsaturated and saturated flow, interception and evaporation, snowmelt and exchanges between aquifers and rivers. Ameli and Creed (2017) and Ameli and Craig (2014) used a 3D groundwater–surface water interaction model to reveal spatially variable recharge rates and groundwater depths. Supported by empirical groundwater table observations, these authors assumed a steady-state subsurface flow.

In the Drava research project, surface connectivity was found to be temporarily limited and of subordinate significance. Therefore, a central task was to search for hydrogeological parameters, which act as boundary conditions for groundwater flow in the floodplain (Dezső et al. 2017). In this approach, the highly changeable water transfer processes can be evaluated in the light of temporally much more stable sedimentological properties (such as grain-size distribution, effective porosity, hydraulic conductivity). Dynamic data on groundwater flow (depth to groundwater table, soil moisture content, rate of seepage from lake) are supplied by groundwater observation wells and soil moisture monitoring.

Table 14.1 Typical effective porosity data of the soil and sediment samples

Sediment	Maximum porosity (V/V%)	Effective porosity (V/V%)
Pale brown loam, root canals	46.52	22.14
Brown, subangular clayey loam	44.78	13.40
Pale brown loamy silt with ferruginous precipitations	43.19	17.15
Grey, single grained, fine sand	48.28	22.90
Pale grey, single grained sand	49.51	30.15

14.3.1 Effective Porosity

Effective porosity (Singhal and Gupta 2010) was measured by the gravimetric method on undisturbed samples. The samples were oven-dried at 105 °C for 24 h and filled with water to saturation. After saturation, samples were placed on a sand bed for 24 h to lose gravitational water.

Although microstratification varied with samples, effective porosity was equally low in all samples. As a consequence, relatively low infiltration rates were found. Effective porosity is reduced by various precipitations (calcareous, ferruginous and organic matter), while increased by root canals and outwash of sediments from the matrix (Table 14.1).

14.3.2 Hydraulic Conductivity

The alluvial sequences in floodplains are rather heterogeneous for physical and hydraulic properties (Wang et al. 2017). The preliminary soil survey, however, allows some typology and a simplified approach can be applied. The sediments of the studied oxbow floor can fundamentally be divided into two types: a clayey-silty and a calcareous sandy unit (Dezső et al. 2017). Hydraulic experiments were carried out on undisturbed sediment samples taken from the deepest part (the former channel thalweg) of the oxbow (Kp1) and parallel with the shoreline (Kp3). The modi of the PSD curves were markedly different, 80 µm along the shoreline and 10 µm in the deepest part of the lake (Fig. 14.2).

Based on hydraulic analyses of the undisturbed sediment samples in the laboratory, different hydraulic conductivity values were found in the middle and shoreline section of the oxbow (from 8.34×10^{-8} to 2.82×10^{-7} m s^{-1})—which is just the opposite to the corresponding pattern in river channels. It is explained by the presence of fine lacustrine silts in the deepest part and the dominance of fine fluvial sands (D_{med} = 80 µm) with high organic matter content in the offshore region.

At the beginning of the hydraulic conductivity experiments, a hydraulic head of 1.5 m was set on the top of the samples (Fig. 14.3). Subsequently, saturated hydraulic conductivity was calculated using the falling head method. The pressure

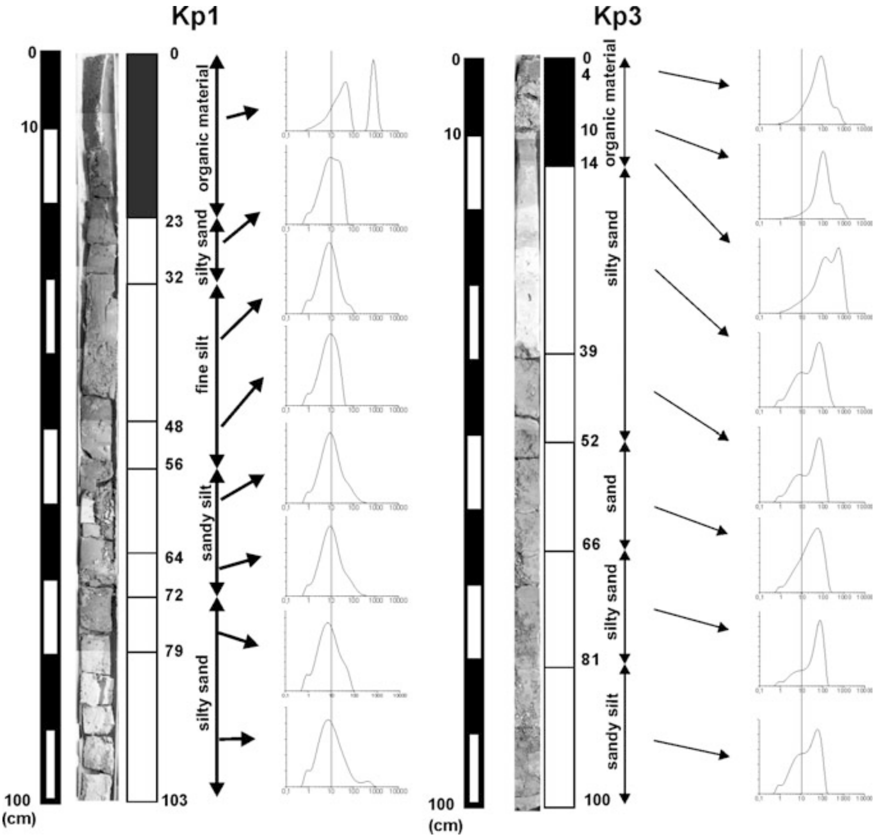

Fig. 14.2 Stratification and grain-size distribution of deposits in the deepest (Kp1) and offshore part (Kp3) of Lake Kisinc (Cún-Szaporca oxbow)

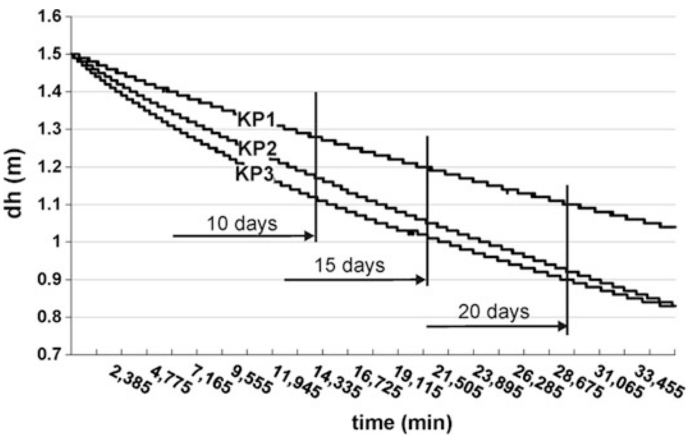

Fig. 14.3 Cumulative infiltration as a function of time for the undisturbed samples Kp1, Kp2 and Kp3

head is representative of the water column height in Lake Kisinc (Fig. 12.3) after the accomplishment of the water replenishment scheme. (Water levels will be raised from the current 90.5 m to an operational level of 91.5–92 m). Saturated hydraulic conductivity was calculated with the following formula (Reynolds and Elrick 1985):

$$k = \frac{L}{(t_2 - t_1)} * \ln\left(\frac{h_1}{h_2}\right)$$

where k is saturated hydraulic conductivity, L is the height of the soil core, t_1 and t_2 are initial and final times of the experiment, respectively, h_1 and h_2 are the corresponding pressure head heights.

Hydraulic conductivity was 8.3×10^{-8} m s^{-1} for Kp1 sediment samples and 2.82×10^{-7} m s^{-1} for the cores Kp2 and Kp3 (Fig. 14.3). When an initial head of 1.5 m was used during the laboratory experiments, water-level drop ranged between 0.38 and 0.60 m for the sediment samples taken from the clogging zones. However, since its cut-off, the oxbow has been functioning as a depositional basin with ever finer sedimentation. The sediments taken from the shoreline borehole originate and have been transported into the lake from the levee of the oxbow. In addition to the dissimilarity in PSD, variations in hydraulic conductivity may be caused by the development of biofilms. (This latter effect, however, cannot be proved from our measurements unambiguously).

14.3.3 Groundwater Flow Modelling During Drought

To study connectivity through groundwater flow, a drought period of 30 days (with no rainfall, 3 mm d^{-1} evaporation) was simulated using the HYDRUS-1D model (Nagy et al. 2017). The model was run for altogether 19 sampled sites with different soil textures under 3 hydrological boundary conditions; dry (ψ < 15,000 H_2O-cm), normal (−15,000 < ψ < −10 H_2O-cm) and saturated conditions or excess ponding (−10 H_2O-cm < ψ).

Multilayered sandy and sandy-silt loam soil profiles showed no capillary rise and groundwater recharge. With groundwater table 1 m below the average depth, permanent wilting point was reached at all sites after a 30-day drought. Water retention varied widely with soil texture type, and the best correlation coefficient (r = 0.79) between observed and measured volumetric water content was found for loam layers in the studied profile.

14.3.4 Seepage from Lake

The rate of seepage from the oxbow can be calculated from the water balance equation. The used input parameters included geomorphological data for the oxbow

Table 14.2 Input and output parameters for seepage calculations

Input data					
From geomorphological survey				From field and laboratory investigations	
Volume of oxbow	V_{oxbow}	(m^3)	Hydraulic conductivity (depending on sedimentological properties of the oxbow and newly flooded areas)	k	(m s^{-1})
Area of oxbow lake surface	A_{oxbow}	(m^2)			
(relative) Water level of oxbow	h_{oxbow}	(m)	Effective porosity	n_{o}	(-)
Output (calculation)					
Amount of seepage from the oxbow				Q_{s}	(m^3 d^{-1})
Total seepage area				A_{s}	(m^2)
Change of oxbow lake surface				A_{om}	(m^2)
Calculated water level of oxbow				h_{cw}	(m a.s.l)
Calculated water storage capacity of oxbow				V_{cws}	(m^3)

and the measured hydrological parameters (Table 14.2). Since the planned replenishment rates for the months of March and June were 43,200 and 24,512 m^3 d^{-1}, respectively (data obtained from DDKÖVÍZIG 2012), a mean 30,000 m^3 d^{-1} replenishment rate was used in our calculations. The thickness of the saturated zone was set to 4 m.

The amount of exchanged water between the oxbow lakes and the surrounding groundwater is proportional to the change of hydraulic head (*dh*), the surface area and the hydraulic conductivity (*k*) of the sediments and inversely proportional to the thickness of the clogging zone (*d*) as it is described by the modified Darcy equation (Brunner et al. 2010):

$$Q_{\text{s}} = \frac{k}{d}\left(h_{\text{ox}} - h_{\text{grw}}\right) * \Delta x \Delta y$$

where Q_{s} is total outflow from the oxbow; k is hydraulic conductivity (m s^{-1}); d is soil depth (m); h_{ox} is relative water level of the oxbow lakes (m); h_{grw} is the depth of the adjacent relative groundwater table (m) and $\Delta x \Delta y$ is the change of seepage surface area that corresponds to A_{s}.

The lakebed was divided into two zones, a shallow zone (less than 1.5 m deep) and a deeper (1.5–2.4 m) zone. The hydraulic conductivities of the recently flooded shoreline areas are similar to the relevant values obtained from the pumping tests. Very different hydraulic conductivity values were found: for the deeper zone (median particle size: $D_{\text{med}} = 10$ μm) an order of magnitude lower hydraulic conductivity ($k \sim 10^{-8}$ m s^{-1}) than in the shallow zone ($D_{\text{med}} = 80$ μm) ($k \sim 10^{-7}$ m s^{-1}). The additionally inundated areas have an even higher conductivity ($k \sim 10^{-5}$ m s^{-1}).

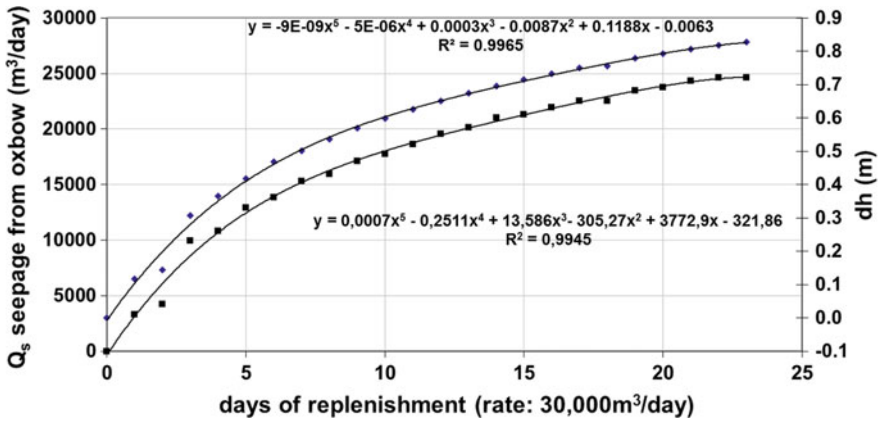

Fig. 14.4 Functional relationship between the hydraulic pressure head difference and seepage

The rising water level in the oxbow triggers an increasing hydraulic pressure difference compared to the adjacent areas. With the increasing volume in the oxbow, the potential contact surface and seepage area also increases. Due to the increasing total seepage area, more and more added water is lost (Fig. 14.4).

14.3.5 Groundwater Flow Modelling with MODFLOW-2005

Water transfer from the oxbow lakes to the main Drava channel was simulated using the MODFLOW 6 model (Langevin et al. 2017). The groundwater flow model includes calculations of groundwater flow (discretization, initial conditions, hydraulic conductance and storage), stress packages (constant heads, wells, recharge, rivers, general head boundaries, drains and evapotranspiration) and advanced stress packages (streamflow routing, lakes, multi-aquifer wells and unsaturated zone flow).

A 10-m resolution Digital Elevation Model (DEM) was built for the area between the Cún–Szaporca oxbow and the Drava main channel. The model is discretized with a finite-difference grid (60 rows, 14 columns and 10 layers; cell size: 10 m × 10 m) (Salem et al. 2017). The eastern and western boundaries are marked with constant head values from previous monitoring of precipitation, groundwater table, infiltration and groundwater recharge. The evapotranspiration is estimated at 13 mm d^{-1}.

The model was run using the replenishment scenarios identified in the Old Drava Programme (DDKÖVÍZIG 2012—see Chap. 21). In scenario 1 (lake level increases by 0.5–91 m above sea level), seepage from the lake was found as 1,298.8 m^3 d^{-1} and water loss through evapotranspiration rise was estimated at

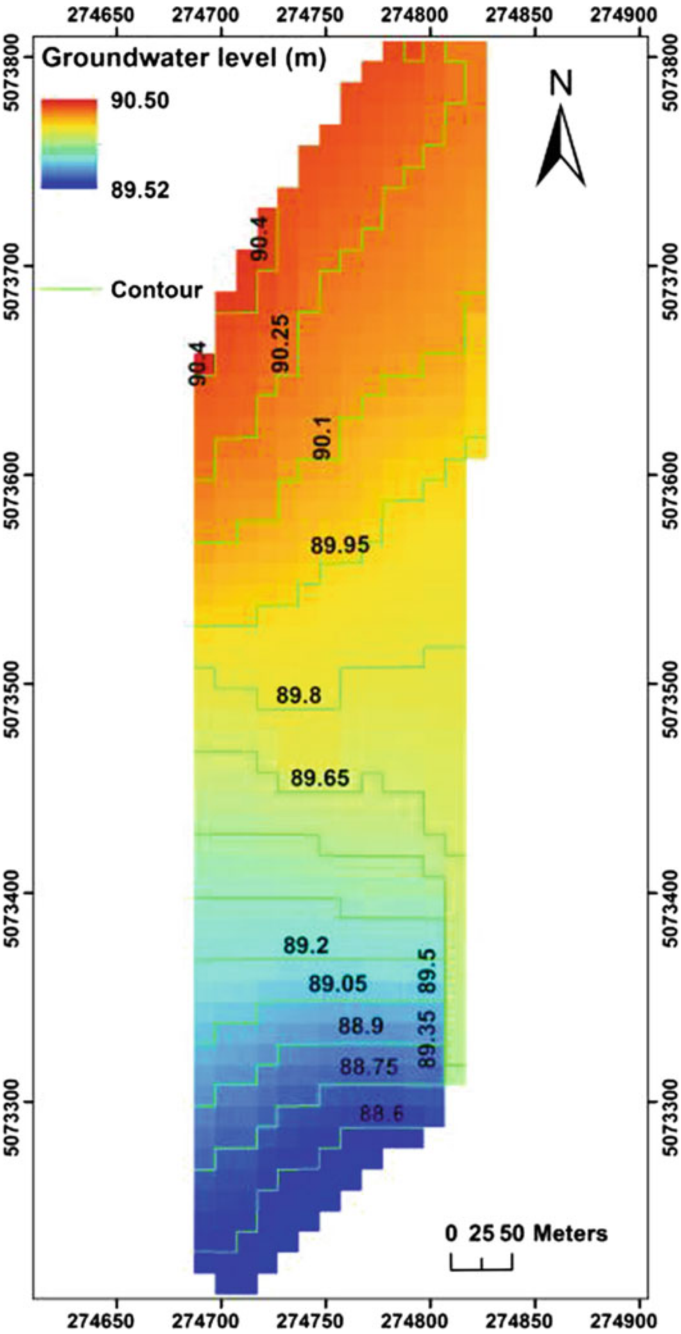

Fig. 14.5 Simulated groundwater levels between the oxbow lakes and the main Drava channel under modelling conditions (by Ali Mohamed Salem)

161.20 m^3 d^{-1} higher than the baseline situation. This equals 0.28 m average groundwater level rise. In scenario 2 (lake level raised by 1 m to 91.5 m), recharge rate from the lake to the aquifer grows by 745.59 m^3 d^{-1}. Consequently, the average water table rises by 0.77 m (Salem et al. 2017).

The advantage of this modelling approach is that it detects water transfer from the lake to inflow into the river. Great variations are revealed according to the grain-size distribution of the alluvial sequence. Calculating with a single uniform aquifer 344.29 m^3 d^{-1} seepage is found for a silty deposit (k = 60 m d^{-1}) and 1,468.95 m^3 d^{-1} for sand (k = 500 m d^{-1}). Recharge to the river is 176.34 m^3 d^{-1} for silt and 1,477.15 m^3 d^{-1} for sand. More realistic estimates of subsurface connectivity are made through the vertical discretization of the aquifer based on the geomorphological interpretation of satellite images supplemented with ground-penetrating radar surveys (Dezső et al. 2017). This way, six layers with different texture and conductivity were distinguished and incorporated in the model (Salem et al. 2017). In this case, water loss amounts to 347.79 m^3 d^{-1} and inflow into the channel is 707.74 m^3 d^{-1}. The figures underline a very high horizontal connectivity between water bodies (Fig. 14.5). At the same time, in the vadose and saturated zones 65% of the water leaked from the oxbow lake may be retained (Salem et al. 2017).

14.4 Conclusions

Surface connectivity of the protected floodplain with the main channel of the Drava is very difficult to provide. Groundwater flow, however, ensures some degree of subsurface hydrological connectivity. This has to be assessed in close association with the water retention capacities of soils and alluvial deposits.

Our research shows that the critical factor in water retention is the transmissivity of lakebed and adjacent deposits. It is not only the present lakeshore that has to be examined hydrodynamically but also the future shallow lakebed zone inundated after water replenishment. Based on the laboratory hydraulic analyses of the undisturbed sediment samples, highly different conductivity values were found for the middle and offshore parts of the oxbow lakes—a pattern just the opposite expected for active river channels. Relatively, coarse fraction ($\sim$80 μm) dominates the shoreline zone and allows higher seepage rate from the oxbow lake. Considerable losses to groundwater (and indirectly to the Drava River) are expected and may jeopardize the success of the replenishment scheme. Allowing for the hydraulic variability of the sediment sequence, HYDRUS-1D and MODFLOW 6 simulations showed to what extent the planned replenishment scenarios will raise the groundwater level and allowed the estimation of soil water retention capacity (Valentová et al. 2010).

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