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ORIGINAL ARTICLE

The spatial variability of streambed vertical hydraulic conductivity in an intermittent river, northwestern China

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Abstract Streambed vertical hydraulic conductivity (*K*) plays an important role in river water and groundwater interaction. The *K* at the ten transects (Ts1–Ts10) at the Donghe River (an intermittent river) in the Ejina Basin, northwestern China, was measured to investigate its spatial variation. Based on the sediment characteristics and vertical hydraulic conductivity of the riverbed, the entire riverbed of the Donghe River could be divided arbitrarily into two parts: an upper part (starting at Ts1 and ending at Ts9, without an obvious and continuous clogging layer) and a lower part (the remaining riverbed, with an obvious and continuous clogging layer). In the upper part, although the *K* varied with depth within the 0–30 cm layer, the variability with depth could be ignored in practice. The arithmetic mean *K* of the upper part ranged from 12 to 27.6 m/day, three orders of magnitude larger than that of the lower part (0.06 m/day). The change of *K* along the river cross section was significant, and larger values of *K* often occurred in the parts of the channels with greater water depth. However, there were no consistent patterns of the variability of *K* at transects across the river, which was influenced by the variation in streambed characteristics. The results could be useful for the estimation of groundwater recharge from river and groundwater resources evaluation in the Ejina Basin.

Keywords Streambed vertical hydraulic conductivity · Ejina Basin · Donghe River · Spatial variability

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Introduction

Streambed hydraulic conductivity plays an important role in surface water and groundwater interaction (Calver 2001; Landon et al. 2001). Determination of the streambed vertical hydraulic conductivity (K) of the entire riverbed has significant importance for the study of groundwater recharge. There are a variety of approaches to determine the hydraulic conductivity of streambed, including numerical modeling (Yager 1993; Sophocleous et al. 1995), environmental tracer experiments (including temperature) (Harvey and Bencala 1993; Hatch et al. 2010; Mutiti and Levy 2010), indoor analysis (Landon et al. 2001; Song et al. 2009; Cheng et al. 2011), and in-stream methods (Chen 2000, 2004; Genereux et al. 2008; Leek et al. 2009; Song et al. 2010). Because of data limitations, numerical modeling and environmental tracer experiments have been less commonly used than indoor analysis and in-stream methods. Indoor analysis refers to the grain-size analysis and indoor determination of the hydraulic conductivity of sample cores (Song et al. 2009; Chen 2011; Cheng et al. 2011). However, grain-size analysis cannot be used to evaluate the influence of sediment structure on hydraulic conductivity (Chen 2000; Kalbus et al. 2006). In-stream methods, including the use of the Standpipe Permeameter, the slug test, and the seepage meter coupled with hydraulic gradient measurement, have been applied widely. The slug test measures only horizontal hydraulic conductivity, although it could be used to determine the permeability of the deeper sediment layer (Landon et al. 2001; Chen 2004; Cheng et al. 2011). The application of the seepage meter is constrained by water depth and flow velocity (Landon et al. 2001; Chen 2004; Kalbus et al. 2006). The Standpipe Permeameter can determine both horizontal and vertical hydraulic conductivity without the limitation of flow velocity (Chen 2000). Moreover, based on a comparison of some in-stream methods, the Standpipe Permeameter has been recommended by Landon et al. (2001) for use in determining the hydraulic conductivity in situ.

The riverbed vertical hydraulic conductivity always varies spatially. Some studies have revealed that the permeability changes significantly along the river cross section (perpendicular to the river flow) (Chen 2005; Genereux et al. 2008; Cheng et al. 2011). Along the river flow (in the downstream direction), even in a small reach (no more than hundreds of meters), the permeability varied remarkably (Chen 2004; Genereux et al. 2008; Hatch et al. 2010). The variation of K with depth has been studied by Song et al. (2010) and Chen (2011). The statistical distribution has also been viewed as an aspect of the spatial variety of streambed hydraulic conductivity (Chen 2005; Genereux et al. 2008; Cheng et al. 2011). For example, Chen (2005) found that the values of streambed K at seven sites out of eight on the Platte River are normally distributed.

The Donghe River is an intermittent river in the Ejina Basin, an extremely arid area (Zhang et al. 2011). The Donghe River is one branch of the lower reaches of the Heihe River which is the second largest inland river in China. Runoff from the Donghe River is the main recharge source of groundwater in the Ejina Basin (Wang et al. 2011a). The runoff of the Heihe River flowing to the lower reaches had been decreased since the 1950s (Wang and Cheng 1998; Si et al. 2009), resulting in many environmental and ecological problems in the Ejina Basin, including a declining water table, vegetation degradation, and aggravated desertification (Feng et al. 2004; Su et al. 2009; Xi et al. 2010; Zhang et al. 2011). In order to protect the environment, the Ecological Water Conveyance Project has been implemented since 2000, and this has restored the groundwater and vegetation to some extent (Zhao et al. 2007; Xi et al. 2010; Wang et al. 2011a, b; Zhang et al. 2011). The permeability of the riverbed is a key factor in the estimation of groundwater recharge from rivers (Chen 2000; Calver 2001; Kelly and Murdoch 2003; Rosenberry and Pitlick 2009). Therefore, determination of the Donghe riverbed hydraulic conductivity is essential to quantify the groundwater recharge in the Ejina Basin.

Several studies have examined the streambed hydraulic conductivity in the Heihe River, especially in the lower reaches. Zhang et al. (2003) measured the saturated hydraulic conductivity of three points in the Donghe River using a Guelph Permeameter (Model 2800k1, Soil Moisture Corporation), but this study ignored the heterogeneity and anisotropy of the sediments. He and Zhao (2007) measured the riverbed permeability in the middle reaches of the Heihe River using a field Standpipe Permeameter, and a preliminary study of the conductivity variation across the river flow has been carried out. Analysis of the spatial variability of riverbed hydraulic conductivity is the basis for studying the variation of river-aquifer system interaction in the lower reaches of the Heihe River. However, as noted above, additional studies are needed to examine the riverbed permeability or the spatial variability of hydraulic conductivity of the lower reaches of the Heihe River. The objectives of this study are to determine the K of the Donghe River using a field Standpipe Permeameter (SP) in situ and an indoor Falling-head Permeameter (FP) in the laboratory and to analyze the variation of K with depth, along the river flow direction and at transects across the river flow.

Study area

The study was conducted at the Donghe River in the Ejina Basin, which is located in Inner Mongolia, northwestern China, between 100°15′–101°15′E and 41°00′–42°16′N (Fig. 1). The Ejina Basin is an extremely arid area with a mean annual precipitation of 36.6 mm and a mean annual temperature of 8.2 °C (Zhang et al. 2011). At Langxinshan, the Heihe River branches into two broad shallow rivers. The Donghe River is the eastern branch and the Xihe River is the western branch, and these rivers ultimately flow into the East and West Juyan Lakes, respectively (Fig. 1). There are two hydrometric stations on the Donghe River: the upper is named the Langxinshan hydrometric station of the Donghe River (LXSD) and the lower is named the East Juyan Lake hydrometric station (EJL) (Fig. 1). Only one hydrometric station is located on the Xihe River, and this is referred to as the Langxinshan hydrometric station of Xihe River (LXSX) (Fig. 1).

The flow was greater before the 1950s, since then the runoff had been decreased (Wang and Cheng 1998). Greater flow rate should lead to greater shear stress and produce a higher permeability layer that consists of coarse sediments. As a result, it is reasonable to speculate that the streambed has at least two typical layers: a fine upper layer and a coarse lower layer. This assumption was verified by a field survey conducted in August 2011, when the riverbed was dry. The upper layer is less permeable than the lower

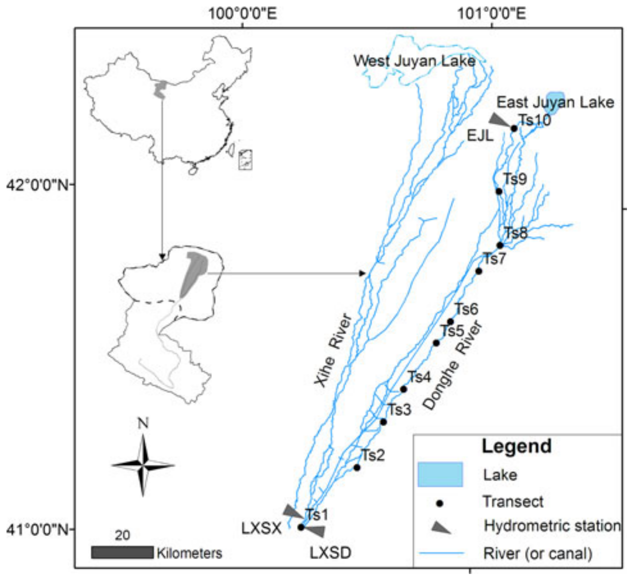

Fig. 1 Location of the study area and the distribution of the measured transects

layer and constrains the hydrologic connectivity between the river and the aquifer. Hence, it is important to measure the hydraulic conductivity of the upper layer when analyzing the river-aquifer system.

Since the Ecological Water Conveyance Project commenced in 2000, the Donghe River has been the main pathway for water delivery (Fig. 1, ESM only). Hence, the Donghe River streambed was selected for the study. Data were collected at ten transects perpendicular to the river channel during the flow period in April 2011 (Fig. 1, ESM only). The ten transects were located between the two hydrometric stations previously mentioned (LXSD and EJL). The ten transects were designated Ts1–Ts10 from LXSD to EJL. Measurements at each transect were decided based on the width of the watercourse. The measured points were designated based on their distance along each transect from the right bank of the Donghe River. That is, the first measurement point at transect Ts2 was designated Ts2-1, and the second measurement point at Ts2 was designated Ts2-2, and so on. The measured points were located at distances from the right bank of the Donghe River from 1 to 450 m.

Methods

Field Standpipe Permeameter test

The field Standpipe Permeameter test (SP) involves inserting a pipe vertically into the streambed, filling the pipe with river water, measuring the rate of decline of the water level, and then calculating the vertical hydraulic conductivity (K_{sp}) using the rate of decline (Fig. 2a). In this study, a polymethyl methacrylate (PMMA) pipe of inner diameter 4.4 cm and length 130 cm was used. The tube was inserted into the streambed sediments, ensuring that the length of the sediment column was approximately 30 cm. River water was poured carefully into the pipe without destroying the surface structure of the streambed. While the initial water head in the pipe (above the river surface) was recorded, the chronoscope was started, which recorded the elapsed time simultaneously. Then, the water head in the pipe was recorded according to the set time interval. Using the water head records at given time intervals, the values of K_{sp} were calculated. During the each test, the water depth was measured at each test location to determine its relationship to streambed hydraulic conductivity.

Hvorslev (1951) used the following equation to calculate K_{sp}:

$$K_{sp} = \frac{\frac{\pi D}{11m} + L_v}{t_2 - t_1} \ln\left(\frac{h_1}{h_2}\right), \tag{1}$$

where D is the inner diameter of the permeameter, m is the square root of the ratio of the horizontal conductivity K_h to the vertical conductivity K_v (i.e., $m = \sqrt{K_h/K_v}$), L_v is the length of the measured sediment column, t_1 and t_2 are the starting and ending times, respectively, and h_1 and h_2 are the corresponding starting and ending water heads, respectively.

Considering that K_h is normally larger than K_v, the modified Hvorslev solution has been suggested by Chen (2000, 2004) to calculate the K_{sp} when L_v is much larger than D:

$$K_{sp} = \frac{L_v}{t_2 - t_1} \ln\left(\frac{h_1}{h_2}\right), \tag{2}$$

However, the error of the solution provided by the modified Hvorslev equation (Eq. (2)) is related to the ratio (L_v/D) of the measured sediment length (L_v) to the inner diameter (D) of the PMMA pipe (Chen 2004). For example, when $1 < m < 5$, if the ratio (L_v/D) is greater than 5, the error of the modified calculation will be less than 5 % (Chen 2004). For this reason, the proper inner diameter and measured sediment column length should be selected carefully. In this study, the average length of the measured sediment column is approximately 30 cm and the inner diameter of the permeameter is 4.4 cm, yielding the ratio of L_v/D greater than 5 and ensuring the precision of these measurements.

A total of 63 measurements were conducted at ten transects in April 2011 (Fig. 1). The location of the ten transects are mapped in Fig. 1, and the number of measurements at each transect is listed in Table 1. K_{sp} was used

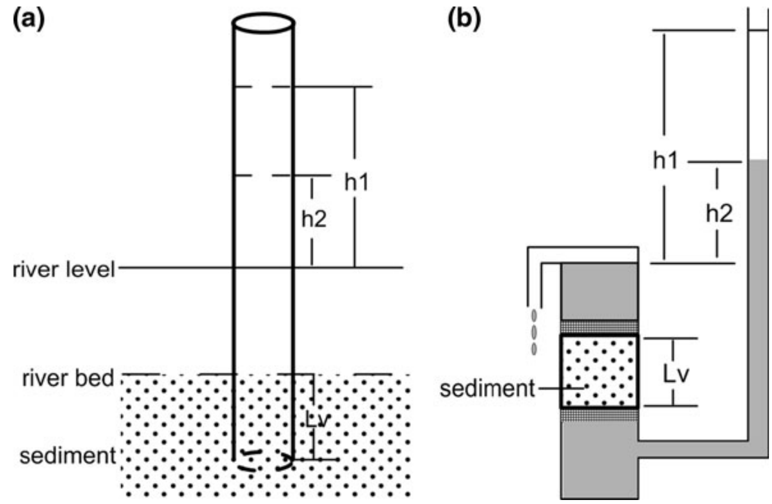

Fig. 2 Schematic diagrams showing: **a** a field Standpipe Permeameter and **b** an indoor Falling-head Permeameter

Table 1 Transect characteristics

Transect name	Riverbed width (m)	Width of the watercourse (m)	Max. water depth (cm)	Min. water depth (cm)	Number of the SP test	Number of the FP test	Riverbed description
Ts1	45	37	45	14	7	4	Fine to Medium sand, without an obvious and continuous clogging layer
Ts2	231	57	36	7.5	6	4	Fine to Medium sand, without an obvious and continuous clogging layer
Ts3	358	58	41	14	6	3	Fine to Medium sand, without an obvious and continuous clogging layer
Ts4	390	43.2	46	10	6	4	Fine to Medium sand, without an obvious and continuous clogging layer
Ts5	440	30	46	7	6	3	Fine to Medium sand, without an obvious and continuous clogging layer
Ts6	336	36	32	11	6	NA	Fine to Medium sand, without an obvious and continuous clogging layer
Ts7	120	70.6	36	10	8	7	Fine to Medium sand, without an obvious and continuous clogging layer
Ts8	134.8	38	33	6	7	7	Fine sand, without an obvious and continuous clogging layer
Ts9	143	23.1	19	7	5	NA	Fine to Medium sand
Ts10	33	30	31	16	6	NA	Fine sand overlying a 1-cm clogging layer

NA no measurements

to illustrate the variation of *K* at transects across and along the river flow.

Sediment sampling and grain-size analysis

The sampling and transportation of undisturbed sediments cores: the sediment cores used for measuring *K* in the laboratory were collected at a depth of ∼0–6 cm and sampled no more than 20 cm from the standpipe test point. The 0–6 cm undisturbed sediments cores were sampled using cylinder core sampler with inner diameter of 4.6 cm and length of 6 cm, made of Polyvinyl Chloride (PVC) pipe. The sampling procedure of sediments core was similar to widely used sampling procedure of undisturbed soil core to some extent. The cylinder core sampler in which the top end was covered by a cap with drain holes was pressed into the riverbed sufficiently without inducing deficiency or compression. Then, an excavating cap was used to excavate the bottom end and the sediments core was carefully removed from riverbed. After inverting the sediments core, superfluous sediments was cleared and another cap with drain holes was used to cover the bottom end. Finally, both the ends of cylinder core sampler with caps were sealed with tape to preserve the sediments structure and to prevent sediments loss. A total of 32 undisturbed sediments cores were sampled, distributed among Ts1–Ts5, Ts7, and Ts8. These sediments cores were placed in a vibration-proof box and were transported carefully to the lab to minimize the disturbance.

The sampling and grain-size analysis of disturbed sediments samples: once the in situ Standpipe Permeameter test was completed, a sediment sample was collected at each test location. By sealing the top opening of the tube of Standpipe Permeameter, the tube could be pulled out with the sediment column inside. The sediment was then packed into a sampling bag. Sediment samples were analyzed using the sieving method and the hydrometer method. The finest cut size was 0.001 mm, and the coarsest cut size was 10 mm. Depending on the texture of the sediment, the sediment samples were separated based on grain size into 6–10 grades. According to a U.S. Department of Agriculture classification standard, particle sizes <0.05 mm are classified as silt and clay, particle sizes of 0.05–2.0 mm are grouped into sand, and particle sizes >2.0 mm are grouped into gravel. The grain-size analysis aided the investigation of the riverbed characteristics.

Indoor Falling-head Permeameter test

The K of the sample cores (K_{fp}) was measured using an indoor Falling-head Permeameter (FP) (Fig. 2b) (Yu et al. 2008) and calculated using Eq. (3)

$$K_{fp} = \frac{aL_v}{A(t_2 - t_1)} \ln\left(\frac{h_1}{h_2}\right), \tag{3}$$

where A and a are the cross-sectional area of sediment core column and the inner cross-sectional area of the pipe, respectively.

The undisturbed sediments cores were carefully loaded into the permeameter; then, the falling-head test was started. The following operational process of the FP test was similar to that of the SP test. The value of K_{fp} was calculated using Eq. (3) and the water head at the given time interval.

Calculation of the value of K of the lower layer

According to Freeze and Cherry (1979), if a horizontal sediment layer consists of two parallel horizontal layers with different vertical hydraulic conductivity, the vertical hydraulic conductivity of the total layer (K) should be calculated using the following equation, Eq. (4):

$$\frac{L}{K} = \frac{L_1}{K_1} + \frac{L_2}{K_2}, \tag{4}$$

where the parameters L, L_1, and L_2 represent the thickness of the total layer, the upper layer, and the lower layer, respectively, and the parameters K, K_1, and K_2 represent the vertical hydraulic conductivity of the total layer, the upper layer, and the lower layer, respectively. In this study, K_{sp} represented K, and K_{fp} represented K_1. K_{cal} was used to represent K for the 6–30 cm layer. Obviously, K_{cal} represented K_2 in the study and could be calculated by the Eq. (4) (Song et al. 2010). K_{fp} and K_{cal} were compared to demonstrate the change of K with depth. It should be mentioned that the measured K values of 0–6 cm sediments cores were not necessarily the true K values of the 0–6 cm sediments in the co-located in situ test because of the heterogeneity of riverbed sediments. This means that the change of K with depth at one measured site may not be well demonstrated by the comparison between K_{fp} and K_{cal}. However, the difference between K values of the 0–6 cm layer (K_{fp}) and K values of the 6–30 cm layer (K_{cal}) could be well illustrated at all the measured sites on the whole.

Statistical analyses

The Shapiro–Wilk Normality Test was used to verify whether the parameters K_{sp} and K_{fp} were normally or log-normally distributed at the 95 % confidence level (Cheng et al. 2011); this was performed using Origin pro 8 software.

Spearman Bivariate Correlation analysis was used to determine whether two non-normal distributed variables are significantly correlated at the 95 % confidence level. The Two-Independent-Samples Test is nonparametric and was used to determine whether or not the differences that existed in two independent groups of K values were statistically significant. The significance of the difference between K_{fp} and K_{cal} was then analyzed using the Two-Independent-Samples Test. The difference between K_{sp} and K_{fp} is considered to be significant when the attained significance level (p value) is less than a predetermined value ($\alpha = 0.05$). These statistical analyses were performed using SPSS 13.0 software.

Results and discussion

The values of K measured using SP and FP

The values of K measured using SP and FP are shown in Table 2. The value of K_{sp} at the ten tested transects ranged

Table 2 Statistics of the measured K_{sp}, K_{fp}, and K_{cal}

Layers	Hydraulic conductivity	Number of measurements	Maximum (m/day)	Minimum (m/day)	Mean (m/day)	Median (m/day)	Standard deviation (m/day)	Coefficient of variation
0–30 cm	K_{sp}	63	53.7	0.02	17.6	16.8	11.8	0.7
0–6 cm	K_{fp}	32	53.1	3.9	19.8	16.4	13.8	0.7
6–30 cm	K_{cal}	32	74.7	6.0	21.0	18.0	16.4	0.8

from 0.02 to 53.7 m/day, and the value of K_{fp} measured at Ts1–Ts5, Ts7, and Ts8 varied from 3.9 to 53.1 m/day; both of these ranges are within the range of K values tabulated by Calver (2001). The tests carried out from Ts1 to Ts9 occurred on a sandy streambed, and the values of K_{sp} varied from ~5.4 to 53.7 m/day. The tests carried out from Ts10 occurred on a sandy streambed that was covered by approximately 1-cm-thick silt and clay deposits, and the values of K_{sp} varied from ~0.02 to 0.1 m/day. Therefore, based on the riverbed sediment characteristics (Table 1) and hydraulic conductivity, the entire riverbed of the Donghe River could be divided arbitrarily into two parts. The upper part is from Ts1 to Ts9 and has no obvious and continuous clogging layer, while the lower part consists of the remainder of the riverbed which extends from Ts9 to the East Juyan Lake and has a clogging layer.

Histograms and Shapiro–Wilk Normality Test results suggested that neither K_{sp} nor ln K_{sp} is normally distributed, a result confirmed at the 95 % confidence level using the Shapiro–Wilk Normality Test ($n = 63$, $p = 0.008$ and $n = 63$, $p = 0.000$, respectively) (Fig. 3). However, K_{fp} is log-normally distributed although not normally distributed at the 95 % confidence level ($n = 32$, $p = 0.21$ and $n = 32$, $p = 0.002$, respectively), as revealed by Shapiro–Wilk Normality Test result.

Although the K values follow commonly observed log-normal distributions (Cardenas and Zlotnik 2003), this viewpoint was not supported by the result of this paper and some other reports (Springer et al. 1999; Ryan and Boufadel 2007; Genereux et al. 2008; Cheng et al. 2011). For example, Genereux et al. (2008) found that streambed K are neither normally nor log-normally distributed in a study of 487 measurements of streambed K values using in situ permeameter tests at West Bear Creek in North Carolina. However, Cheng et al. (2011) reported that K was normally distributed in the Platte River. The result implied that the lognormal distribution of K should be used discreetly in this study area.

The value of K varied with depth

The parameters K_{fp} and K_{cal} represented the K of the upper 0–6 cm and lower 6–30 cm layers, respectively. The value of K_{cal} varied from 6.0 to 74.7 m/day among transects Ts1–Ts5 and Ts7, with an arithmetic mean value larger than the arithmetic mean of K_{fp}. The variability of K with depth was illustrated by comparing K_{fp} and K_{cal}. The two groups of K values are shown in Fig. 4, and the K_{fp} values were larger than the K_{cal} values at some test points. For example, the K_{fp} value could be 3.5 times larger than K_{cal} at transects Ts7-8 (Fig. 4). At other test points, such as Ts2-2 and Ts8-7, the K_{fp} values were less than the K_{sp} values. At Ts8-7, the value of K_{cal} value was ~18 times larger than that of K_{fp}, and this result could be caused by the local and discontinuous clogging layer. Based on the correlation analysis, K_{fp} and K_{cal} are not significantly correlated at the 95 % confidence level ($n = 32$, $r = 0.012$, and $p = 0.948$). The Two-Independent-Samples Test revealed that there is no significant difference between K_{fp} and K_{cal} at the 95 % confidence level ($n = 32$, $p = 0.83$), indicating that the difference between the K-value of the upper 0–6 cm sediment layer and that of the lower 6–30 cm sediment layer was not generally statistically significant. Therefore, the comparison indicated that the value of K varied with depth within the 0–30 cm sediment layer and there is neither a statistically significant correlation nor a statistically significant difference between the K values of the upper and lower layers.

The variation of K along cross section

Figure 5 shows the variation of K_{sp} at transects across the river flow. It can be seen that the value of K_{sp} decreased as the distance to the right bank increased in transects Ts1, Ts2, Ts6, Ts7, and Ts9. In contrast, the value of K_{sp} increased as the distance to the right bank increased in transects Ts3, Ts4, Ts5, and Ts10. The values of K_{sp} at the other two transects (Ts8 and Ts9) appeared not to vary with distance from the river bank. The K_{sp} of Ts7 varied the most, from 5.0 to 53.7 m/day. There was no consistent pattern in the variation of K_{sp} with distance to the right riverbank throughout the Donghe River. Variations of K across the river cross section have been widely reported (Landon et al. 2001; Chen 2004, 2005; Genereux et al. 2008). For example, Chen (2004) found that the value of K varied from 0.8 to 77.8 m/day in the Platte River streambed. Landon et al. (2001) and Chen (2005) reported that the variation could occur over short distances such as

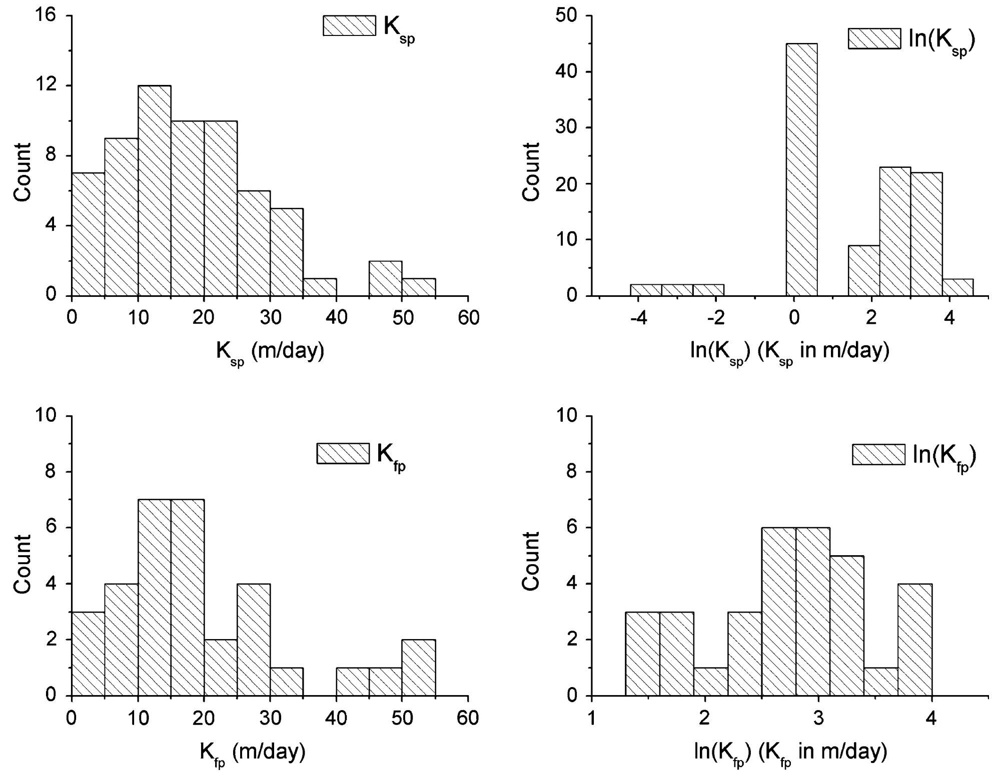

Fig. 3 Histogram of K_{sp} and K_{fp}

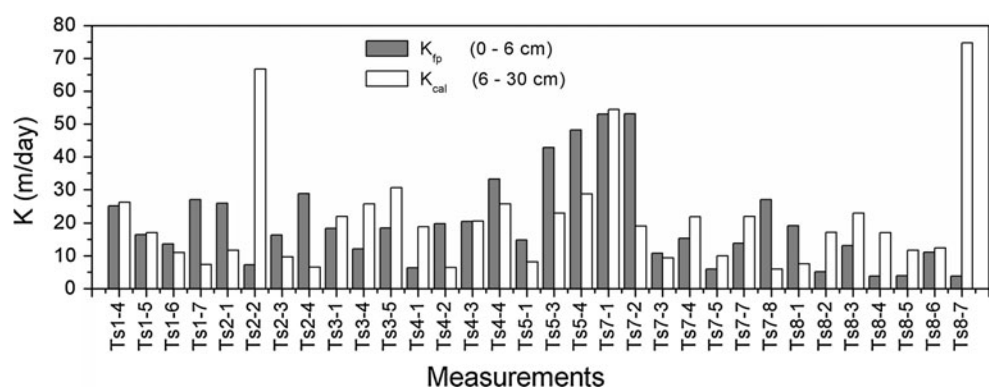

Fig. 4 Difference between the 0–6 cm layer and the 6–30 cm layer illustrated by comparing K_{fp} and K_{cal} at the measurement points

3 m. However, although the variation of K_{sp} was inconsistent at different transects in this study, Genereux et al. (2008) reported that the K value at the center of the channel was usually higher than the K value close to the riverbank at West Bear Creek. In any case, the transverse variation of K demonstrated that the values of K should be measured at different locations across cross sections of the Donghe River for the results to be representative.

Downstream changes of K

The arithmetic mean of K_{sp} varied from Ts1 to Ts9 (in the upper part of the river that did not have an obvious and continuous clogging layer), ranging from 12 to 27.6 m/day; in contrast, the arithmetic mean of K_{sp} at Ts10 (in the lower part of the river that has a continuous clogging layer) was approximately only 0.06 m/day (Fig. 6). It should be noted that the clogging layer has a strong influence on hydraulic conductivity, as verified by Landon et al. (2001) and Chen (2004) in the Platte and Wood Rivers, respectively. In the lower reaches of Heihe River, the K value has also been determined by other researchers. Wu et al. (2005) obtained a value of K for the entire Donghe Riverbed as $\sim$5–17 m/day using a numerical method; this value is almost consistent with in situ test results in this study, with the exception of the K value measured at Ts10. Using a Guelph

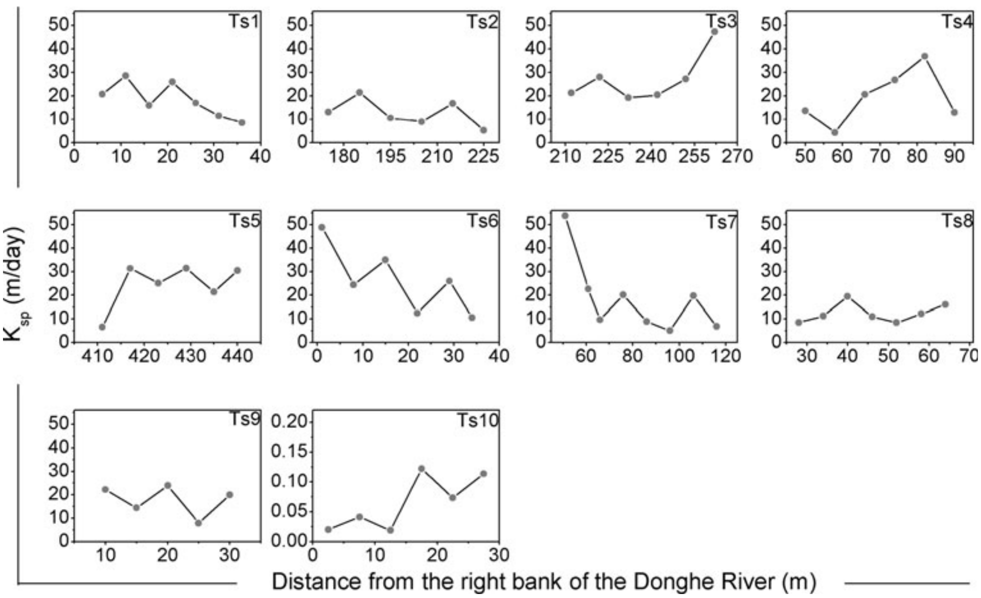

Fig. 5 The variation of K_{sp} along each transect measured using a field Standpipe Permeameter

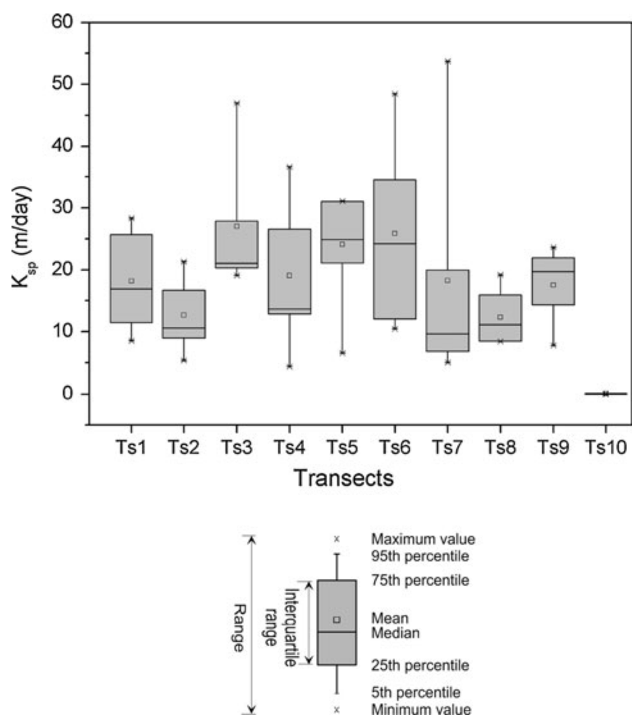

Fig. 6 Box plots of streambed K_{sp} at ten transects in the Donghe River

Permeameter, Zhang et al. (2003) measured the value of K of the Donghe River bed in the reaches between Ts1 and Ts9, and determined K values in the range 0.3–0.9 m/day. However, the reason for the difference between the results of this study and that of Zhang et al. (2003) is unclear. Although the K values obtained in the various studies are not the same, the magnitude of the K values obtained in this study does not differ greatly from that obtained in the previous studies of this area.

The dependence of K on water depth and streambed characteristics

The vertical hydraulic conductivity of riverbeds should be affected by the grain-size distribution of its sediments and the sedimentary structure, both of which are strongly related to the environment in which sedimentation took place (Leek et al. 2009). In general, the more rapid the flow velocity, the deeper the water depth and the larger the flow strength, both of which could result in coarser sediments. Therefore, the water depth was considered to reflect flow strength and hence might be used to explain the variation in K_{sp}. Correlation analysis was used to determine whether the water depth affected K_{sp} in transects Ts1–Ts9. Although the results revealed that the correlation is not significant at the 95 % confidence level ($R = 0.185$, $n = 57$, $p = 0.169$), the value of K_{sp} increased with water depth (Fig. 7). However, the positive relationship between K_{sp} and water depth indicated that larger values of K often occurred in the deeper parts of the channels.

The spatial distribution of K was related to regional variations in streambed characteristics (Landon et al. 2001; Leek et al. 2009; Rosenberry and Pitlick 2009) including sedimentary texture and sedimentary structure; these characteristics are controlled by the depositional and erosional processes that are influenced by the temporally and spatially varied river flow in rivers in arid regions (Walters 1990; Tooth 2000). The flow of the Donghe River is controlled artificially with high temporal variation (Fig. 1, ESM only). Most of the river water recharges the groundwater by transmission losses (Xi et al. 2010; Wang et al. 2011a; Zhang et al. 2011), which leads to spatial variations in the river flow. The depositional and erosional processes under varying spatial and temporal river flows

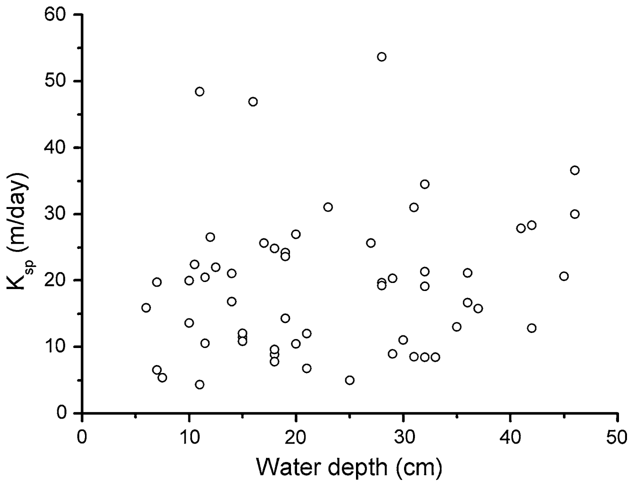

Fig. 7 Relationship between water depth and K_{sp} in the upper reaches of the river

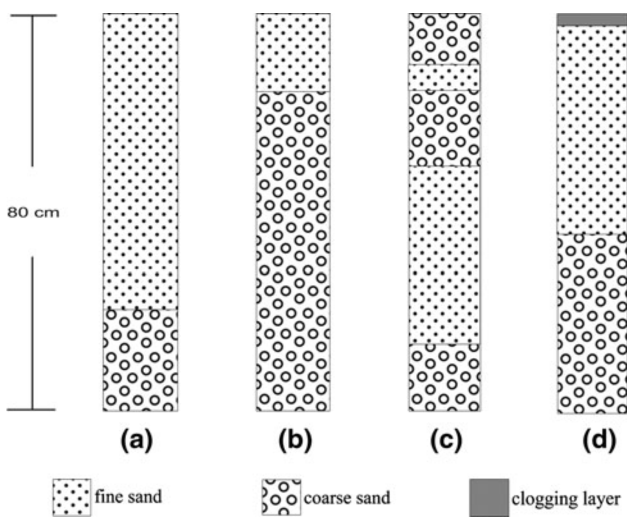

Fig. 8 Schematic diagrams of four typical patterns of sedimentary structure based on the field survey of August 2011: **a**, **b**, **c**, and **d** are located at Ts1, Ts4, Ts7, and Ts10, respectively

lead to a strong spatial variation of streambed texture and sedimentary structure, ultimately causing the hydraulic conductivity of the Donghe River to vary.

The grain-size distribution varied significantly at different locations, even at the same river cross section (Fig. 2, ESM only), and this could result in the variability of *K*. In addition, not only did the grain-size distribution vary, but also did the sedimentary structure. In general, in a sedimentary structure, the lower layer is more compact than the upper layer due to the pressure caused by the upper layer; consequently, the conductivity of the lower layer is lower than that of the upper layer (Song et al. 2010). However, according to our field survey, the grain size of the upper layer in the Donghe Riverbed is generally finer than that of the lower layer, although heterogeneity exists, particularly in the 0–30 cm layer. Furthermore, based on the field survey, at least four types of sedimentary structure were found in the Donghe riverbed, as shown in Fig. 8: (a) a thick fine sand layer (~70 cm) overlies a coarse sand layer; (b) a thin fine sand layer (~10 cm) overlies a coarse sand layer; (c) fine and coarse sand layer alternately; and (d) a fine sand layer (~40 cm) underlies a 1-cm-thick clogging layer and overlies a coarse sand layer. Moreover, even at the same transect, the sedimentary structure is heterogeneous at different measurement sites.

The values of K_{fp} and K_{cal}, which represent the *K* of the 0–6 cm and the 6–30 cm sediment layers respectively, differed due to the sediment characteristics and sedimentary structure. K_{cal} is greater than K_{fp} if the measured site is located at an area with the second or the fourth types of sedimentary structure (e.g., Ts2-2, Ts4-1, and Ts8-7), whereas the two *K* values have no apparent difference for areas with the first type (e.g., Ts1-4, Ts1-5, and Ts4-3); K_{cal} is smaller than K_{fp} for areas with the third type (e.g., Ts4-2, Ts7-2, and Ts7-8). This result suggests that heterogeneity exists in the 0–30 cm sediment layer. However, based on the Two-Independent-Samples Test, in transects Ts1–Ts9, where the riverbed was not covered by an obvious clogging layer, the variability of *K* with depth in the 0–30 cm layer could be ignored in practice.

Conclusions

In this paper, the streambed vertical hydraulic conductivity (*K*) of the Donghe River in the Ejina Basin was measured at ten transects. The value of *K* varied with depth, along the river flow direction and at transects across the river flow.

- Based on the riverbed sediment characteristics and the vertical hydraulic conductivity, the entire riverbed of the Donghe River could be divided arbitrarily into two parts: an upper part that did not have an obvious and continuous clogging layer (starting at Ts1 and ending at Ts9) and a lower part that did have an obvious and continuous clogging layer (the remaining riverbed).
- The *K* values of the Donghe riverbed covered a wide range, varying from ~0.02 to 53.7 m/d. In the upper part of the river, the arithmetic mean of the *K* value ranged from ~12 to 27.6 m/day. However, in the lower part of the river, the arithmetic mean of the *K* value was only 0.06 m/day.
- The value of *K* varied with depth within the 0–30 cm layer affected by sedimentary structure. Nevertheless, the variability of *K* in the upper part of the river could be ignored in practice, whereas the variability of *K* with depth in the lower part of the river should be emphasized.
- The change of *K* at transects perpendicular to the river flow was significant and larger values of *K* often occurred in the deeper part of channels. The patterns of

variation at transects across the river flow were multiply affected by the versatile riverbed sediment characteristics of the Donghe River bed.

The results of this work should be useful for modeling river-aquifer systems and understanding the recharge of groundwater by surface runoff in such a remote and intermittent river. In order to obtain more information regarding groundwater recharge in the region, the values of K of deeper sediment layers of the riverbed should be measured or simulated in future studies.

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References

Calver A (2001) Riverbed permeabilities: information from pooled data. Ground Water 39(4):546–553. doi:10.1111/j.1745-6584.2001.tb02343.x

Cardenas MB, Zlotnik VA (2003) Three-dimensional model of modern channel bend deposits. Water Resour Res 39(6):1141. doi:10.1029/2002wr001383

Chen XH (2000) Measurement of streambed hydraulic conductivity and its anisotropy. Environ Geol 39(12):1317–1324

Chen XH (2004) Streambed hydraulic conductivity for rivers in south-central Nebraska. J Am Water Resour Assoc 40(3):561–573. doi:10.1111/j.1752-1688.2004.tb04443.x

Chen XH (2005) Statistical and geostatistical features of streambed hydraulic conductivities in the Platte River, Nebraska. Environ Geol 48(6):693–701. doi:10.1007/s00254-005-0007-1

Chen XH (2011) Depth-dependent hydraulic conductivity distribution patterns of a streambed. Hydrol Process 25(2):278–287. doi:10.1002/hyp.7844

Cheng C, Song JX, Chen XH, Wang DM (2011) Statistical distribution of streambed vertical hydraulic conductivity along the Platte River, Nebraska. Water Resour Manag 25(1):265–285. doi:10.1007/s11269-010-9698-5

Feng Q, Liu W, Su YH, Zhang YW, Si JH (2004) Distribution and evolution of water chemistry in Heihe River Basin. Environ Geol 45:947–956. doi:10.1007/s00254-003-0950-7

Freeze RA, Cherry JA (1979) Groundwater. Prentice-Hall, Englewood Cliffs

Genereux DP, Leahy S, Mitasova H, Kennedy CD, Corbett DR (2008) Spatial and temporal variability of streambed hydraulic conductivity in West Bear Creek, North Carolina, USA. J Hydrol 358(3–4):332–353. doi:10.1016/j.jhydrol.2008.06.017

Harvey JW, Bencala KE (1993) The Effect of streambed topography on surface-subsurface water exchange in mountain catchments. Water Resour Res 29(1):89–98. doi:10.1029/92wr01960

Hatch CE, Fisher AT, Ruehl CR, Stemler G (2010) Spatial and temporal variations in streambed hydraulic conductivity quantified with time-series thermal methods. J Hydrol 389(3–4):276–288. doi:10.1016/j.jhydrol.2010.05.046

He ZB, Zhao WZ (2007) Measurement of streambed hydraulic conductivity and anisotropy analysis. Adv Water Sci 18(3):351–355 (in Chinese)

Hvorslev MJ (1951) Time lag and soil permeability in ground-water observations. Waterways Experiment Station, Corps of Engineers, US Army Bulletin 36, Vicksburg, Mississippi

Kalbus E, Reinstorf F, Schirmer M (2006) Measuring methods for groundwater–surface water interactions: a review. Hydrol Earth Syst Sci 10(6):873–887. doi:10.5194/hess-10-873-2006

Kelly SE, Murdoch LC (2003) Measuring the hydraulic, conductivity of shallow submerged sediments. Ground Water 41(4):431–439

Landon MK, Rus DL, Harvey FE (2001) Comparison of instream methods for measuring hydraulic conductivity in sandy streambeds. Ground Water 39(6):870–885. doi:10.1111/j.1745-6584.2001.tb02475.x

Leek R, Wu JQ, Wang L, Hanrahan TP, Barbet ME, Qiu HX (2009) Heterogeneous characteristics of streambed saturated hydraulic conductivity of the Touchet River, south eastern Washington, USA. Hydrol Process 23(8):1236–1246. doi:10.1002/hyp.7258

Mutiti S, Levy J (2010) Using temperature modeling to investigate the temporal variability of riverbed hydraulic conductivity during storm events. J Hydrol 388(3–4):321–334. doi:10.1016/j.jhydrol.2010.05.011

Rosenberry DO, Pitlick J (2009) Local-scale variability of seepage and hydraulic conductivity in a shallow gravel-bed river. Hydrol Process 23(23):3306–3318. doi:10.1002/hyp.7433

Ryan RJ, Boufadel MC (2007) Evaluation of streambed hydraulic conductivity heterogeneity in an urban watershed. Stoch Environ Res Risk Assess 21(4):309–316. doi:10.1007/s00477-006-0066-1

Si JH, Feng Q, Wen XH, Su YH, Xi HY, Chang ZQ (2009) Major ion chemistry of groundwater in the extreme arid region northwest China. Environ Geol 57(5):1079–1087. doi:10.1007/s00254-008-1394-x

Song JX, Chen XH, Cheng C, Wang DM, Lackey S, Xu ZX (2009) Feasibility of grain-size analysis methods for determination of vertical hydraulic conductivity of streambeds. J Hydrol 375(3–4):428–437. doi:10.1016/j.jhydrol.2009.06.043

Song JX, Chen XH, Cheng C, Wang DM, Wang WK (2010) Variability of streambed vertical hydraulic conductivity with depth along the Elkhorn River, Nebraska, USA. Chin Sci Bull 55(10):992–999. doi:10.1007/s11434-009-0640-2

Sophocleous M, Koussis A, Martin JL, Perkins SP (1995) Evaluation of simplified stream-aquifer depletion models for water rights administration. Ground Water 33(4):579–588. doi:10.1111/j.1745-6584.1995.tb00313.x

Springer AE, Petroutson WD, Semmens BA (1999) Spatial and temporal variability of hydraulic conductivity in active reattachment bars of the Colorado River, Grand Canyon. Ground Water 37(3):338–344. doi:10.1111/j.1745-6584.1999.tb01109.x

Su YH, Zhu GF, Feng Q, Li ZZ, Zhang FP (2009) Environmental isotopic and hydrochemical study of groundwater in the Ejina Basin, northwest China. Environ Geol 58(3):601–614. doi:10.1007/s00254-008-1534-3

Tooth S (2000) Process, form and change in dryland rivers: a review of recent research. Earth Sci Rev 51(1–4):67–107. doi:10.1016/s0012-8252(00)00014-3

Walters MO (1990) Transmission losses in arid region. J Hydraul Eng ASCE 116(1):129–138. doi:10.1061/0733-9429

Wang GX, Cheng GD (1998) Changes of hydrology and ecological environment during late 50 years in Heihe River Basin. J Desert Res 18(3):233–238 (in Chinese)

Wang P, Yu JJ, Zhang YC, Fu GB, Min LL, Ao F (2011a) Impacts of environmental flow controls on the water table and groundwater chemistry in the Ejina Delta, northwestern China. Environ Earth Sci 64(1):15–24. doi:10.1007/s12665-010-0811-0

Wang P, Zhang YC, Yu JJ, Fu GB, Ao F (2011b) Vegetation dynamics induced by groundwater fluctuations in the lower Heihe River Basin, northwestern China. J Plant Ecol 4(1–2): 77–90. doi:10.1093/jpe/rtr002

Wu Q, Xu JX, Zhang ZZ, Ma ZM (2005) Coupled modeling of surface water–groundwater system II. Application. Shu Li Xue Bao 36(6):754–758 (in Chinese)

Xi HY, Feng Q, Liu W, Si JH, Chang ZQ, Su YH (2010) The research of groundwater flow model in Ejina Basin, northwestern China. Environ Earth Sci 60(5):953–963. doi:10.1007/s12665-009-0231-1

Yager RM (1993) Estimation of hydraulic conductivity of a riverbed and aquifer system on the Susquehanna River in Broome County. U.S. Geological Survey Water-Supply Paper 2387, New York

Yu ZB, Huang Y, Schwartz FW (2008) Ground water hydrology. Science Press, Beijing (in Chinese)

Zhang YH, Wu YQ, Qiao MY (2003) Lysimetric experiment on the riverbed in the lower reaches of Heihe River. Arid Zone Res 20(4):257–260 (in Chinese)

Zhang YC, Yu JJ, Wang P, Fu GB (2011) Vegetation responses to integrated water management in the Ejina basin, northwest China. Hydrol Process 25(22):3448–3461. doi:10.1002/hyp.8073

Zhao WZ, Chang XL, He ZB, Zhang ZH (2007) Study on vegetation ecological water requirement in Ejina Oasis. Sci China Ser D 50(1):121–129. doi:10.1007/s11430-007-2035-z