ground water

Dynamics of Flood Water Infiltration and Ground Water Recharge in Hyperarid Desert

by Ofer Dahan[1], Boaz Tatarsky[2], Yehouda Enzel[3], Christoph Kulls[4], Mary Seely[5], and Gererdo Benito[6]

Abstract

A study on flood water infiltration and ground water recharge of a shallow alluvial aquifer was conducted in the hyperarid section of the Kuiseb River, Namibia. The study site was selected to represent a typical desert ephemeral river. An instrumental setup allowed, for the first time, continuous monitoring of infiltration during a flood event through the channel bed and the entire vadose zone. The monitoring system included flexible time domain reflectometry probes that were designed to measure the temporal variation in vadose zone water content and instruments to concurrently measure the levels of flood and ground water. A sequence of five individual floods was monitored during the rainy season in early summer 2006. These newly generated data served to elucidate the dynamics of flood water infiltration. Each flood initiated an infiltration event which was expressed in wetting of the vadose zone followed by a measurable rise in the water table. The data enabled a direct calculation of the infiltration fluxes by various independent methods. The floods varied in their stages, peaks, and initial water contents. However, all floods produced very similar flux rates, suggesting that the recharge rates are less affected by the flood stages but rather controlled by flow duration and available aquifer storage under it. Large floods flood the stream channel terraces and promote the larger transmission losses. These, however, make only a negligible contribution to the recharge of the ground water. It is the flood duration within the active streambed, which may increase with flood magnitude that is important to the recharge process.

Introduction

Scarcity of rain and lack of streamflow make ground water of shallow alluvial aquifers the most important source of water in arid and hyperarid environments. This ground water is used daily in many parts of the world's deserts. More importantly, this water resource may buffer the impact of dry seasons and even of relatively longer droughts on riparian ecosystems, wildlife, and human settlements. The ability to exploit ground water in arid regions controls the livelihood of communities, whether they are small rural ones or larger ones practicing modern intensive agriculture or industry.

Ground water recharge in desert environments is controlled by two main mechanisms: (1) direct regional infiltration of rain water in the mountains and interdrainage areas and (2) flood water infiltration through ephemeral channel beds (also known as transmission loss) (Osterkamp et al. 1994; Schwartz 2001; Shentsis and Rosenthal 2003; Walter et al. 2000). Direct infiltration in arid environments is relatively ineffective because of the rarity of rainstorms, low mean average precipitation and high potential evaporation. Therefore, in many desert areas, direct rain infiltration is regarded as nonexistent (Scanlon 2004). Furthermore, the high potential evaporation relative to precipitation in arid environments results in soil salinity and the rare deep infiltration of rain water reduces water quality due to salinization. Therefore, ground water in most of the world's deserts is

[1]Corresponding author: Zuckerberg Institute for Water Research (ZIWR), J. Blaustein Institutes for Desert Research, Ben-Gurion University of the Negev, Sede Boqer Campus 84990, Israel; (972) 8-6596917; fax: (972) 8-6596909; odahan@bgu.ac.il
[2]Zuckerberg Institute for Water Research (ZIWR), J. Blaustein Institutes for Desert Research, Ben-Gurion University of the Negev, Sede Boqer Campus 84990, Israel.
[3]Institute of Earth Sciences, The Hebrew University of Jerusalem, Givat Ram Campus, Jerusalem 91905, Israel.
[4]Institut Für Hydrologie, Albert-Ludwigsuniversität, Freiburg 79085, Germany.
[5]Desert Research Foundation of Namibia, Windhoek, Namibia.
[6]Centro de Ciencias Medioambientales, Madrid 28006, Spain.
Received October 2007, accepted November 2007.

doi: 10.1111/j.1745-6584.2007.00414.x

OnlineEarlyArticles
doi: 10.1111/j.1745-6584.2007

characterized by relatively high salinity (Simmers 1997). Flood water infiltration through beds of ephemeral channels, also known as transmission loss, relies on short-duration water flow in stream channels (Shentsis and Rosenthal 2003). The transmission loss partially recharges the local alluvial aquifers underneath stream channels and the aquifers connected to them. In contrast to the regional recharge by rainfall, flood water is usually characterized by low salinity since the streambed and its alluvium are more regularly flushed by such floods (Scanlon 2004). Therefore, in arid environments, ground water of acceptable quality usually lies in the shallow alluvial aquifers along the stream channels (de Vries and Simmers 2002; Gee and Hillel 1988). Accordingly, water resource management in arid environments must rely on quantifying this flood water infiltration and percolation which recharge the shallow alluvial aquifers.

Several methods are traditionally used to estimate flood water infiltration and ground water recharge. These include (1) *transmission losses*—calculating the water mass balance based on the volume difference between two gauging stations along a stream channel and interpreting the recharge rate from the water loss (Lange 2005; Shentsis et al. 1999); (2) *ground water level variation*—wells located in the vicinity of the channel allow interpretation of the flood water infiltration rate through the observed or modeled increase in the aquifer's water storage (Blasch et al. 2004; Sanford 2002); and (3) *chemical and isotopic composition*—flood water composition is compared with the composition of the various water bodies surrounding and influencing the ground water in the vicinity of the stream channel (Scanlon 2004). The latter method allows an evaluation of the proportion of each water source in the aquifer and consequently the recharge component. Note that in hyperarid environments, data that can support the use of the previously mentioned methods is very scarce and therefore their application is limited.

All these methods are widely used in flood water infiltration studies. However, direct measurements of the processes leading to recharge from channel bed to ground water do not exist. Therefore, flood water infiltration and ground water recharge are usually evaluated from parameters measured at the system boundaries. Furthermore, quantification of the actual hydrological processes taking place in the vadose zone is rarely attempted because of a lack of monitoring methodologies. Therefore, hydrological process in the vadose zone is usually left unmonitored and consequently unknown.

This paper presents new results from a study on flood water percolation through the vadose zone and the ground water recharge during natural flood events in ephemeral rivers in a hyperarid environment. We perform direct measurements and quantifications of the rates of infiltration, percolation, and recharge of shallow alluvial aquifers. We present field measurements made by a monitoring system installed in the bed of the lower Kuiseb River in the Namib Desert, western Namibia. The monitoring system continuously recorded (1) the flood stages above the channel bed; (2) the water content variation along the entire vadose zone profile below the stream channel; and (3) the changes in ground water levels. The long-term goal is to link the flood hydrology to the percolation process, which will provide new insight into long-term quantification of ground water recharge in arid environments.

Study Area

The Kuiseb River is one of the largest ephemeral rivers (~560 km) in Namibia, draining the western Great Escarpment of western Namibia with a 420-km long catchment area of approximately 15,500 km^2 (Jacobson et al. 1995). It drains the high plateau (~2000 m in elevation) westward through the escarpment into the Atlantic Ocean near Walvis Bay (Figure 1). The middle and lower reaches of the Kuiseb River cut their way through the central Namib Desert, forming the boundary between the vast sand dune sea to the south and the gravel flat peneplains to the north (Scholz 1972). Mean annual rainfall is greater than 300 mm/year in the headwater and decreases to less than 20 mm/year in the lowlands. Mean annual potential evaporation is 1700 to 2500 mm/year, increasing from the coast of the Atlantic Ocean inland (Botes et al. 2003).

The Gobabeb area (Figure 1) is underlain by metamorphic rocks of the late Proterozoic Damara Group. The Salem Granites that intruded the Damara Group are exposed in the lower Kuiseb River and they likely form the boundaries of the Kuiseb River's shallow alluvial aquifer around Gobabeb. This aquifer supplies the water for the Gobabeb Research and Training Center and the nearby villages. The thickness and width of the alluvial fill of the riverbed vary. It is composed mainly of light-colored, slightly micaceous, well-sorted quartz sand (Botes et al. 2003).

The relatively wetter headwater of the Kuiseb River can generate a flood almost every year that overcomes the transmission losses along its long course and attains its lower hyperarid reaches. In the past 20 years (except 2001), the river has flowed pass Gobabeb at least once a year for an average of 12 d/year, with a maximum of 33 d in 1997. Similar to other rivers with wet headwater and hyperarid lower reaches (Bueno and Lang 1980; Enzel 1990), only rare floods are capable of flooding the entire length of the Kuiseb River: in only approximately 10% of the years (over the past 120 years) have the Kuiseb River floods been large enough to overcome transmission losses along its lower course and reach the Atlantic Ocean near Walvis Bay.

Hydrologically, the Kuiseb water resources are one of the two most diversely used among the ephemeral rivers in Namibia. Along the entire course of the river, surface runoff and ground water are exploited for drinking, farming, and mining, with the main consumer being the city of Walvis Bay, situated at the river mouth on the coast of the Atlantic Ocean. In total, a population of more than 30,000 people with tens of thousands of tourists annually depend directly on the water of the Kuiseb River (Botes et al. 2003).

A geophysical study at the study site mapped the boundaries of the alluvial aquifer. The width of the active stream channel along the Kuiseb River varies from 25 to

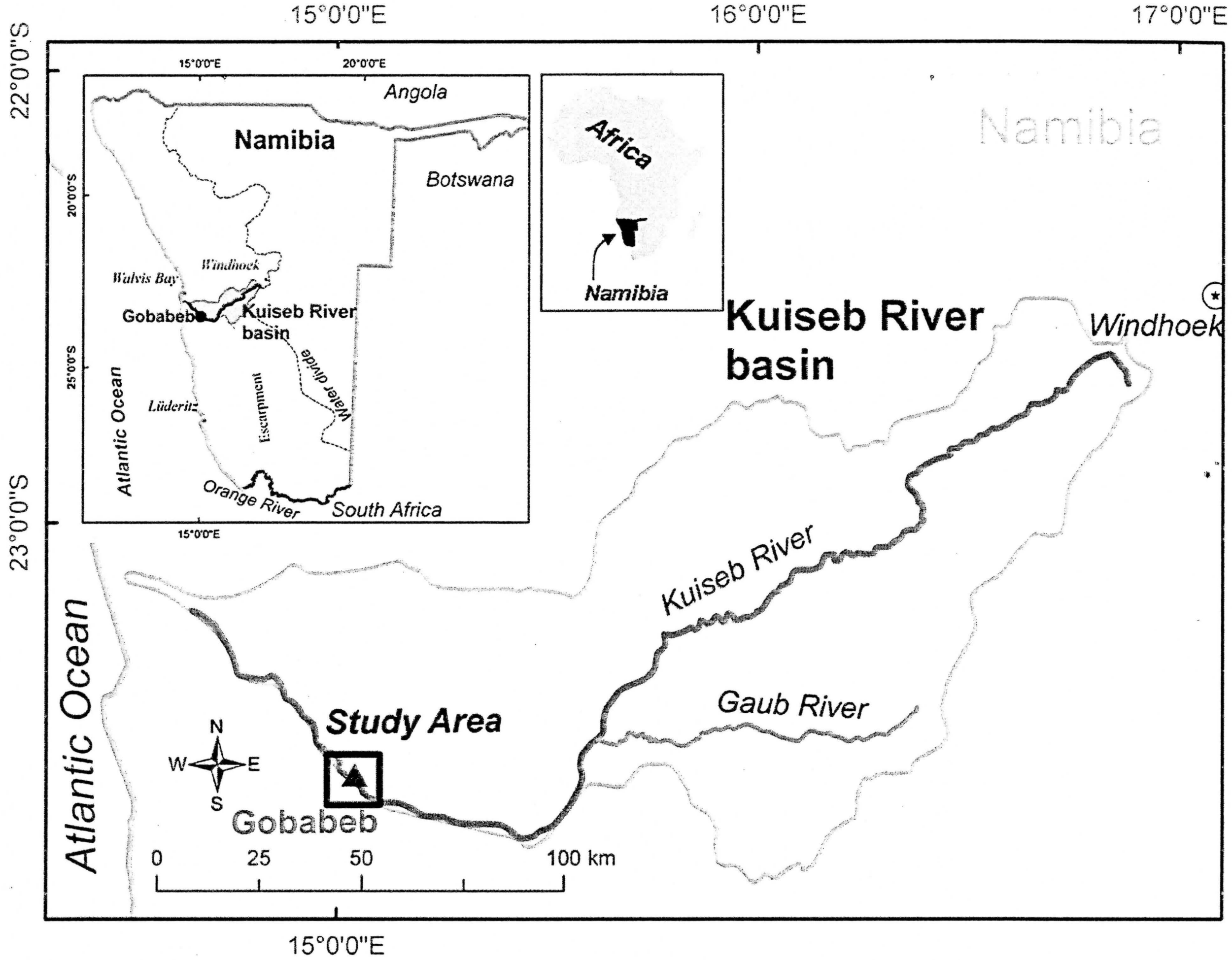

Figure 1. The Kuiseb River catchment and location of the study site near Gobabeb, Namibia.

60 m. Figure 2 is a typical cross section of the channel showing a width of 33 m at the study site. The active channel is relatively flat (Figure 2), with riparian trees such as *Faidherbia albida, Acacia erioloba*, and *Tamarix usneoides* growing in the flood plains (Jacobson et al. 1995). The channel bed and most of the flood plains are composed of relatively well-sorted sandy sediments (Table 1).

Methods

Concepts

The study implemented a monitoring setup (Figure 3) that simultaneously measures the flood stage, the water content variation in the vadose zone, and the ground water response to the recharge process. A detailed description of the monitoring system and its performance is provided in previous publications (Dahan et al. 2003, 2006; Dahan et al. 2007; Rimon et al. 2007). Nevertheless, for the sake of convenience, a short description of the system is presented here. The monitoring system allows the installation of multilevel Flexible Time Domain Reflectometry (FTDR) probes at any desired depth with minimum disturbance of the soil column above the probes. Consequently, continuous, real-time measurement of water content variations at multiple depths throughout the entire vadose zone is possible. The monitoring system was implemented in a small-diameter (15 cm), slanted (35°) borehole, drilled especially for the installation of the FTDR probes (Figure 4). The probes are made of flexible stainless steel waveguides attached to a flexible sleeve, made of a thin polyvinyl chloride liner. The flexible sleeve with the FTDR probes on its upper side is then inserted into the slanted borehole and filled with liquid two-component urethane. The large hydrostatic pressure generated inside the sleeve by the liquid resin pushes the probes very tightly against the borehole walls, ensuring full contact with the surrounding soil. Consequently, the entire borehole volume is packed with the sleeve which follows the borehole wall's roughness and diameter irregularities. A slanted drilling method was selected to ensure monitoring of an undisturbed sedimentary column. Assuming that the general flow direction in the vadose zone is vertical, each point on the upper side of the slanted borehole faces an undisturbed sediment column. The FTDR system is operated with Campbell Scientific data acquisition and logging instruments including TDR100, SDM50X, and AM16/32 multiplexers, and CR10X data loggers. All waveguides are connected to the acquisition

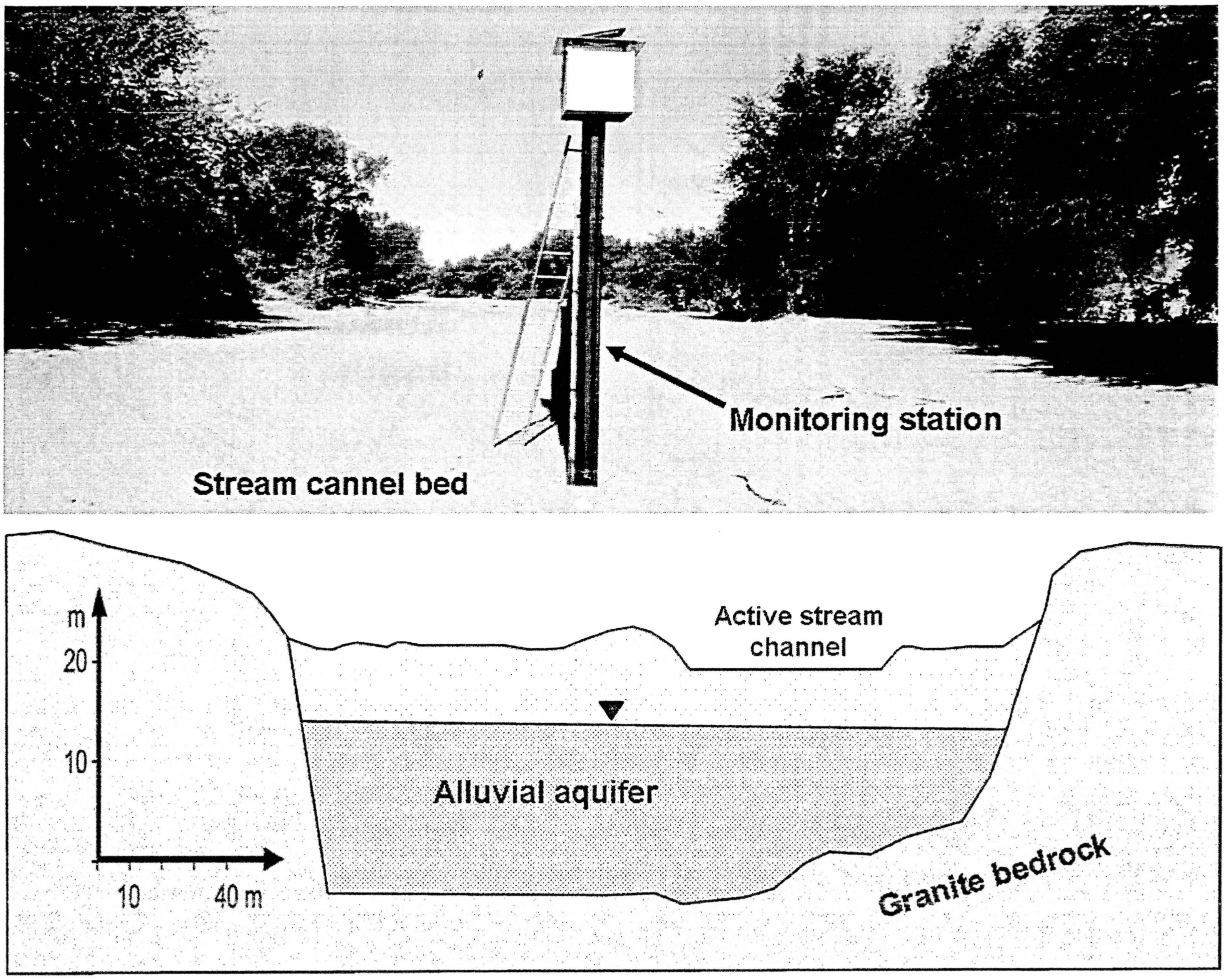

Figure 2. The Kuiseb River stream channel and the monitoring station by the Gobabeb research centre (above) and a typical lithological cross section of the Kuiseb River aquifer at the study area (below).

system using a high-quality RG-58 coax cable (shielding factor greater than 95%). Specific calibration curves for soil moisture and FTDR probes are presented elsewhere (Dahan et al. 2003; Rimon et al. 2007).

Ground water levels were monitored using a pressure transducer (Levelloger M5, Silinst Canada Ltd.) placed in a 4-inch piezometer that was installed in the stream channel with perforated screens to the upper 5 m of the ground water. The piezometer was installed near the vadose zone monitoring station (Figure 4). At the surface, the piezometer was extended inside a protection tower that was also used to host the data logging equipment and the flood-stage gauging system. Installation of the monitoring system was completed in July 2005.

Results and Discussion

The Kuiseb River usually runs no more than once or twice a year. Luckily, the summer rainy season of January to March 2006 was exceptionally wet and the river experienced five flood events that passed through Gobabeb within a short period (Figure 5a). The floods ranged from small ones lasting only 2 d with peak flood stages of 0.5 m to relatively large ones that flowed for 15 d with relatively high peak stages that exceeded 3 m. During that period, the water content variation along the vadose zone as well as water table fluctuations were measured simultaneously (Figures 5b and 5c). Figure 5 shows how each flood in the stream channel initiates an infiltration process in the vadose zone which ultimately leads to ground water

Table 1
Grain-Size Distribution in Samples from the Unsaturated Zone at the Gobabeb Monitoring Station

Depth (m)	0.6	1	1.5	2	2.75	3	4	6
Gravel (%)	0.01	0.00	0.12	8.82	51.80	4.87	0.67	0.46
Sand (%)	99.43	99.35	99.18	90.76	47.06	93.30	97.14	92.09
Silt + clay (%)	0.56	0.65	0.70	0.42	1.14	1.83	2.19	7.44

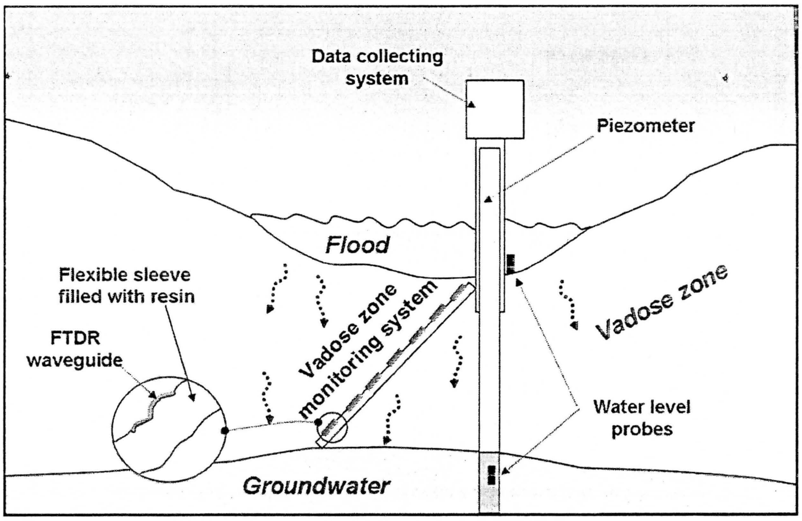

Figure 3. Schematic illustration of the methodological and monitoring system (not to scale).

recharge. This is expressed as a sequential increase in vadose zone water content and, ultimately, as a distinct rise of the water table. The exceptional sequence of floods with different characteristics enabled a detailed evaluation of flood water infiltration and ground water recharge. Detailed data and analysis from the first and second floods will be presented first, showing the infiltration process through the synchronous reaction of the vadose zone and ground water to the flood. Then, the integrated results from the entire season will be described to elucidate the general relationships between the flood and stream-channel characteristics and the ground water recharge process.

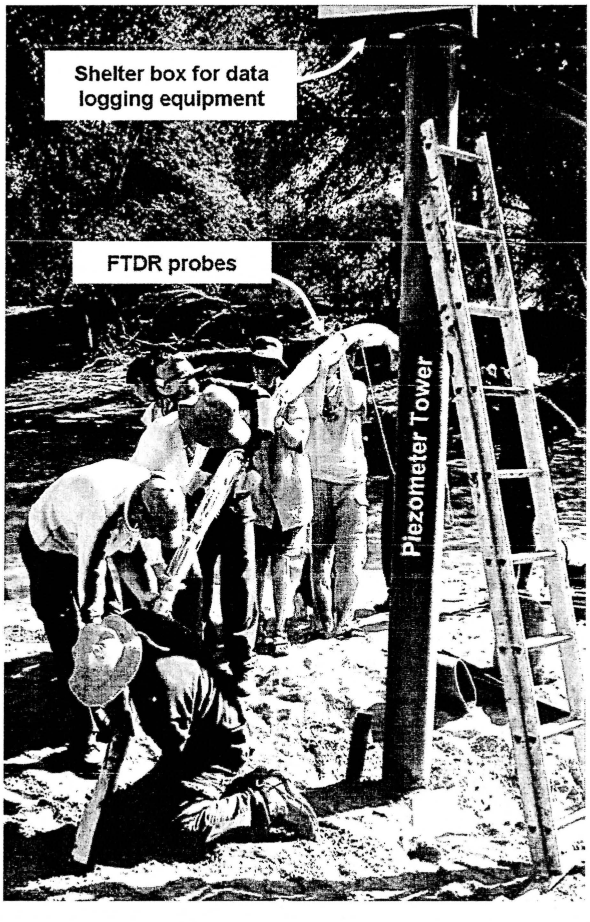

Figure 4. Installation of the FTDR probes in a slanted borehole beneath the stream channel.

Percolation

The data collected in the field during the floods were analyzed following a conceptual model which describes several main stages in the infiltration process (Figure 6). These include (1) schematic illustration of subsurface cross section; (2) propagation of a wetting front from the stream channel down through the vadose zone toward the ground water; (3) a rise in water table due to the recharge process; and (4) water level relaxation and stabilization on the new water level by end of the recharge event.

The first flood was recorded by the monitoring station on January 20, 2006. The river flowed for 76 h (~3 d) and its maximum water level reached approximately 1.5 m (Figure 7a). Downward water percolation was recorded by the FTDR probes soon after the flood's arrival.

Figure 7b presents the water content variation in the vadose zone during and after the first flood. The first sharp increase in water content marks the arrival of the wetting front to each depth. This stage is described in phase b of the conceptual model (Figure 6b). The wetting front seems to propagate uniformly along the entire vadose zone, and the probes are sequentially wetted from the first probe at 0.58 m (probe 1) to the deepest probe at 4.59 m (probe 7).

The water content variation in the sediments during percolation is a function of the flow conditions above and below each layer, as well as the physical characteristics of the layer itself, such as porosity and grain-size distribution. While the flow conditions above and below each layer control inflow and outflow, the porosity and grain-size distribution control the layer's water retention, field capacity, and saturation degree. The change in the sediments water content (Figure 7b) appears to correspond

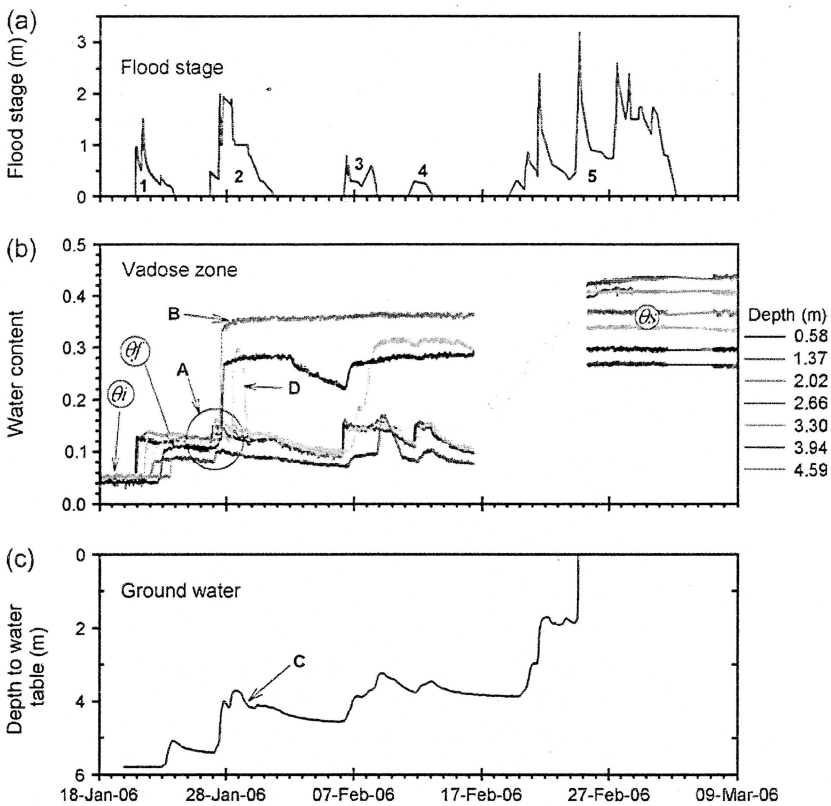

Figure 5. Hydrographs of (a) flood stage in the stream channel, (b) water content variation at various depths along the vadose zone under the stream channel, and (c) ground water levels measured in observation well located in the stream channel during the January to March 2006 flood season. Floods are numbered from 1 to 5, and the symbols *θi*, *θf*, and *θ*s denote initial, final, and saturated water content values, respectively.

well with the lithological composition of the sediments in the monitored area (Table 1). It clearly shows that the smallest increase in water content appears at approximately 2.7 m in coarser sediments, which obviously has low water retention. The initial water content, which represents the residual water content in the vadose zone 11 months after the previous flood, was measured as approximately 5% for the entire profile (Figure 7b). Although the river was flowing continuously for more than 3 d, the water content along the entire depth of the vadose zone, including the upper probes, remained unsaturated with relatively low water content, ranging between 8% and 14%.

The variation in ground water level in response to the first flood event shows that the water table started to rise more or less when the wetting front was detected in the deepest probe (Figure 7c). This rise indicates an increase in ground water storage due to the percolation process observed in the vadose zone. This stage is represented in phase c of the model (Figure 6c). As soon as the flood ceased, the infiltration process stopped, the vadose zone partly drained, and the ground water level started to decrease toward a new water level, which was higher than the initial one (Figure 6d and B in Figure 7c).

The second flood reached Gobabeb monitoring station on January 26. The river flowed for 120 h (5 d) and the maximum flood level reached 2 m. The infiltration process was detected through the water content variations in the vadose zone and the rise in the ground water (Figure 8). However, in this flood, the observed infiltration process seemed quite different from the first flood. Figure 8 is an expansion of A in Figure 5b. Although initial water contents for the second flood were 8% to 12%, higher than the initial value (5%) of the first flood (Figures 7b and 8), an increase in water content was monitored by the FTDR probes immediately after the flood. Close examination of the probe-wetting sequence shows that the wetting front propagated in an orderly fashion downward from the first probe at 0.58 m to the deepest probe at approximately 4.59 m (the water content rises from 12% to 15%). However, the average increase in vadose water content in this flood was only 3.3%, compared with 6.7% during the first flood. A closer look shows that the water content started to increase downward all the way to the deepest probe, but then an abrupt and dramatic increase in water content was observed in probe 7 (stage A in Figure 8 and point B in Figure 5b). This time, the water content reached saturation levels of 35%. The dramatic increase in water content to saturation is attributed to the water table rising to the probe level, shown schematically in Figure 6c (phase c of the conceptual model). The timing of probe 7's saturation corresponds exactly to the time at which the ground water rose to that probe's depth (Figures 5b and 8). The same process of orderly wetting from the ground water up into the vadose zone was recorded by probe 6 and later by probe 5 (stages B and C in Figure 8). During this flood, the

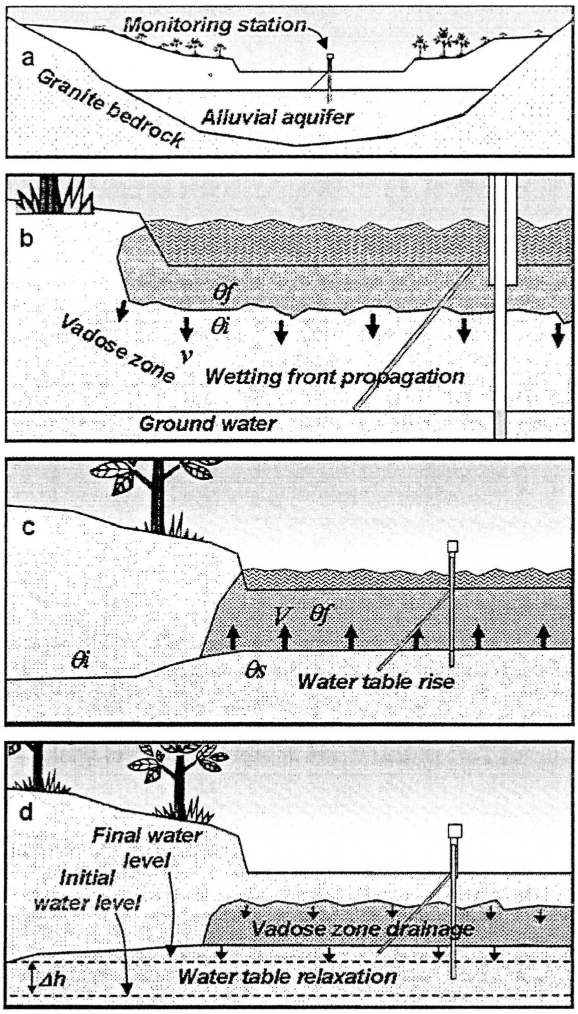

Figure 6. Conceptual phases in flood water infiltration and ground water recharge: (a) typical cross section of an alluvial aquifer in an arid region (similar to the cross section of the study area), (b) early stages of infiltration where a wetting wave propagates from the streambed down through the vadose zone toward the ground water, (c) wetting front arrives at the ground water and water table rises up toward the vadose zone, and (d) infiltration stops, the vadose zone is drained, and the water table stabilizes at a higher level.

ground water rose up to 3.7 m below the surface, bringing the probe at depth of 3.3 m to saturation. However, the water content at this probe dropped sharply from saturation (~30%) to field capacity (13%) (D in Figure 5b), with ground water relaxation (C in Figure 5c and Figure 6d). Similar to the first flood, the entire vadose thickness remained at relatively low water content and did not reach saturation although the river was flooded for 5 d.

Flow Velocity and Fluxes

Vadose Zone

The wetting sequence of the FTDR probes allows direct calculation of the wetting-front propagation velocity v

$$v = \Delta z / \Delta t \tag{1}$$

where Δz is the vertical distance between two adjacent probes and Δt is the time gap between their responses to a change in water content. The response time was defined as the time at which the probe reached 50% of the total increase in water content. Combining the calculated wetting-front propagation velocity (v) with the measured change in water content $\Delta\theta$ allowed calculation of the downward flux q as follows:

$$q = v \times \Delta\theta \tag{2}$$

where $\Delta\theta$ is defined as the difference between the initial water content (θ_i) and final water content (θ_f) (Figure 7). Table 2 is a summary of the calculated wetting-front propagation velocities and fluxes for the first and second floods. During the second flood, the wetting response time is shorter and the calculated velocity of wetting propagation is twice the measured velocity during the first flood. However, the change in water content during the second flood is half of the change measured during the first flood. As a result, very similar average downward fluxes in the vadose zone (~1 cm/h) were calculated for both floods.

Ground Water

The fluctuation in water table levels during and after a flood reflects the dynamic relationship between the influx from the vadose zone to the ground water and the ground water's lateral flow. The following characteristics of the stream channel and aquifer structure enable a calculation of the recharging fluxes directly from the data on water table changes: (1) the active stream channel is flat and much wider (33 m) than the shallow depth (~5 m) to the ground water; (2) the alluvial aquifer is only approximately 200 m wide as it is bounded by the granite bedrock (Figure 2) from both sides; and (3) our sampling and analyses indicate that the channel bed material is composed of relatively homogeneous sand (based on the data presented in Table 1 and additional five boreholes which were drilled at the study site and are not presented here). Assuming that the flood infiltrates across the entire width of the river, we can associate the rise in water table level with an increase in ground water storage. Moreover, since the measuring point is in the middle of the channel, away from the boundaries of the recharging zone, the rate of change in water level during the early stages of the recharge will not be much different from the recharge rate. Accordingly, the rising limb of the ground water hydrograph (A in Figure 7c) reflects the maximum rate of recharge. Hence, the maximum flux q

$$q = V(\theta s - \theta f) \tag{3}$$

where θf is the water content of the unsaturated sediments under the stream channel during the flood event, θs is the saturated water content (Figures 5b, 6c, and 7b), and V is the rate of change in water level:

$$V = \Delta h / \Delta t \tag{4}$$

where Δh and Δt are measures on the rising limb of the hydrograph. The calculated fluxes for the first and second floods are 1.4 and 1.38 cm/h, respectively (Table 2). Note

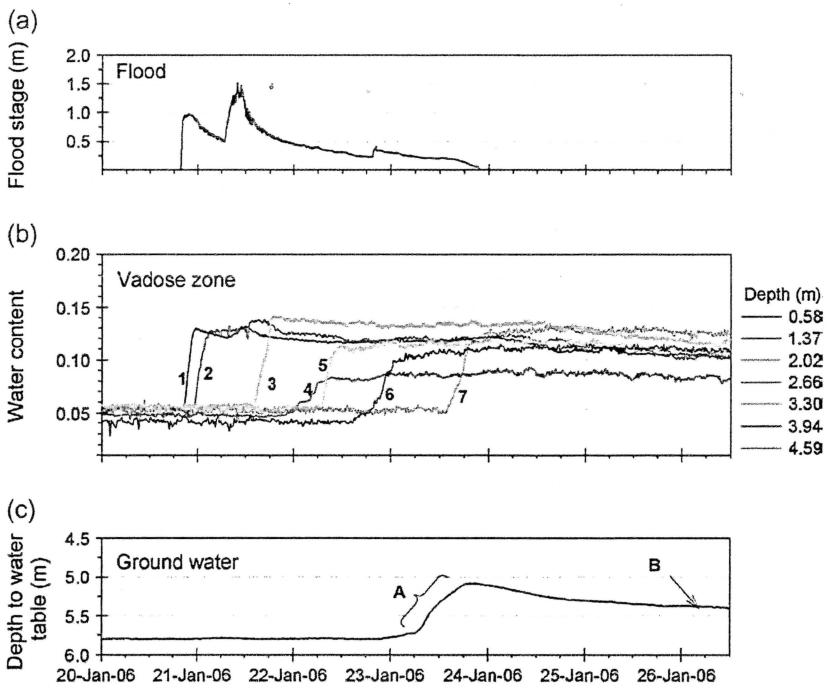

Figure 7. Hydrographs of (a) flood stage in the stream channel, (b) water content variation at various depths along the vadose zone under the stream channel, and (c) ground water levels measured in an observation well located in the stream channel during the first flood. This is an expansion of the data from Figure 5. The symbols θi and θf denote initial and final water content values measured before and after each flood.

that while the fluxes measured in the vadose zone by the wetting-front propagation velocity and the variation in water content were carried out during the early stages of the percolation process. This method calculates the fluxes during later times and represents the maximum rate. Nevertheless, the calculated fluxes are quite similar for both floods and for both methods.

The fluctuations in ground water level may also be used to calculate the average flux for the entire recharge event through the total increase in ground water storage. In this method, the flux is calculated from the difference between the initial water level before the flood and the final water level after relaxation (Δh) when the water level stabilizes at a higher level than the initial one (B in Figure 7c and Figure 6d). At this stabilization phase, the water level presents the full change in ground water storage over the entire width of the alluvial aquifer (Figures 6a and 6d). Accordingly, the total storage increase per unit length of the aquifer Q_l for an entire flood event is calculated as follows:

$$Q_l = \Delta h[(\theta s - \theta f)b + (\theta s - \theta i)(l - b)] \qquad (5)$$

where Δh is the total change in water level after relaxation, $(\theta s - \theta f)$ is the water content increase in the sediments under the stream channel, b is the stream channel

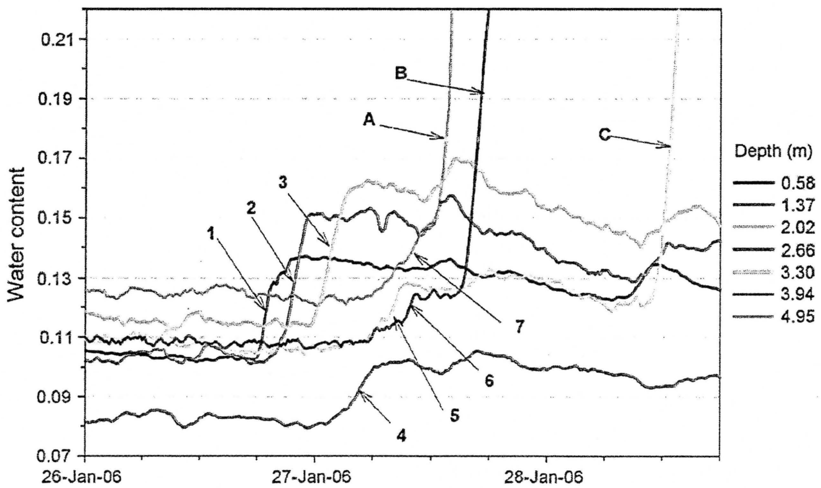

Figure 8. Water content variation in the vadose zone during the second flood. This is an expansion of the circled area A in Figure 5b.

Table 2
Calculated Fluxes During the First Two Floods

		Flood 1, January 20, 2006 at 20:00			Flood 2, January 26, 2006 at 16:45		
Method	**Depth (m)**	**$\Delta\theta$ (%)**	**v (cm/h)**	**q (cm/h)**	**$\Delta\theta$ (%)**	**v (cm/h)**	**q (cm/h)**
Vadose zone	0.58	7.5	33.1	2.5	3.8	29.0	1.1
	1.37	7.8	28.7	2.2	5.0	28.7	1.4
	2.02	8.2	4.0	0.3	4.8	16.3	0.8
	2.66	3.8	5.4	0.2	2.3	23.3	0.5
	3.3	5.8	16.0	0.9	2.5	23.3	0.6
	3.94	7	4.7	0.3	1.6	42.7	0.7
	4.59	7.7	3.3	0.3			
	Average	6.7	13.6	1.0	3.3	27.2	0.9
Ground water	Water level rising rate			1.4			1.38
	Increase in ground water storage			0.7			0.90

width, $(\theta s - \theta i)$ is the water content increase in the sediments under the flooding terraces, and l is the aquifer width (~200 m). Attributing the value of the total recharge per unit length to the active stream channel width (~33 m) and the total flood duration t results in an average recharge flux:

$$q = Q_l/(t \times b) \quad (6)$$

The resultant calculated fluxes are 0.7 and 0.9 cm/h for the first and second flood, respectively. Again, the fluxes calculated from the total increase in ground water storage are very similar to those calculated from the vadose zone and water table data, although the calculations are based on different methods (Table 2).

Flood Characteristics and the Recharge Process

Wetting of the vadose zone by each flood proceeded in an orderly fashion; it progressed from the surface to the water table, followed by a water level rise up into the vadose zone toward land surface. For as long as the wetting progressed downward, the water content in the vadose zone did not exceed field capacity values and remained at 10% to 15%. The upper part of the vadose zone remained unsaturated although the surface was flooded for several days. When the wetting front reached the water table and ground water started to rise up toward the unsaturated zone, a secondary wetting process progressing from the water table up into the vadose zone began. In this process, the sediments reached saturation (27% to 42%). By the second peak of the fifth flood, the entire vadose zone was saturated (Figure 5); the flood flowed directly on a fully saturated aquifer, and for all practical purposes, water infiltration ceased at this river reach. Along its long route, the Kuiseb River usually transmits all of its flood water to the channel bed and it rarely flows all the way to the ocean. During the fifth flood, on January 25, ground water reached the surface (Figure 5), and the river could not lose any additional water by transmission to the vadose zone as the storage capacity of the aquifer was full. Interestingly, at this exact date, the river flowed all the way to the Atlantic Ocean. Such an event is very rare and requires the significant reduction in transmission loss capacity that was provided by ground water rise to the surface.

The calculation procedure described in detail for the first and second floods was then applied to the three additional floods of the 2006 season (Figure 5). Unfortunately, due to a technical problem, some of the FTDR data during the early stages of the fifth flood has been lost. Therefore, the analysis of the fifth flood relay only on ground water data. Table 3 summarizes the calculated fluxes from all floods. Although the floods were very different from one another, the calculated fluxes do not vary much, ranging between 0.4 and 1.5 cm/h. This suggests that the fluxes are limited by a flux-regulating mechanism at the top of the vadose zone. This may be the sediment texture and/or structure, which regulates the downward

Table 3
Calculated Recharge Fluxes for All Floods by Various Methods

	Flux (cm/h)				
Calculation Method	**First Flood**	**Second Flood**	**Third Flood**	**Fourth Flood**	**Fifth Food**
Wetting-front propagation in the vadose zone	1	0.9	1.5	0.7	
Water table rising rate	1.4	1.38	0.57	0.39	1.05
Increase in ground water storage	0.71	0.91	0.83		

water percolation. Closer inspection of the sedimentary structure indicates a very finely layered (laminae) structure on the millimeter scale with alternating coarse and fine sand and minor silt content. These are not distinguishable clay or silt layers. Analysis of grain-size distribution in these samples would categorize them as sand. Since alluvial formations are created by flood deposits, it is expected that these formations would always consist of alternating layers, even if they are composed mainly of sand. We propose that the alternation of layers acts as the regulator that limits infiltration rates and buffers the fluxes, even during the relatively high flood stage. An additional mechanism that may act as a flux-regulating mechanism could be related to clogging of the sediment pores by fine particles from the flowing flood (Blasch et al. 2004). This, however, is a trivial and constant process that should happen with any flood, including those that create the alluvial formation itself. Accordingly, it may be argued that a new flood adds additional clogging particles which were not added in previous floods.

One of the most significant phenomena observed in the Kuiseb River during the 2006 flood season was that small floods yield recharge fluxes very similar to those generated by large floods. This, however, should be regarded with caution since this observation contradicts that from a previous work which suggested that the infiltration under very low flood stages may be limited. Dahan et al. (2007) attributed the limited infiltration rates into dry sediments of a stream channel bed at low flood stages to physical phenomena such as sediments, hydrophobicity, and capillary barriers. It is possible, however, that those limiting factors are not active when infiltration takes place under prewetted conditions. Lange (2005) reported that small floods in the Kuiseb River may flow longer distances due to reduced transmission losses. They attributed the reduced transmission losses to thin silt and clay deposits which may be found on the stream channel bed after flood events and suggested that only large flood events are capable of removing those layers and thus allow significant transmission losses. However, it should be noted that most floods arrive when the layer is cracked due to desiccation and easily detached from the underlying sand. Accordingly, this layer is probably removed from the surface by the first flood bore. The removal of this layer by flood water is supported by the observation that the silt deposits on the active channel bed always overlie clear sand. Moreover, similar silt layers were never observed in the stratigraphy of the vadose zone in auger drillholes and pits dug into the sediments underlying the channel bed in the study area; that is, they are never buried by sand in the active channel and the layers are solely a surface feature. In addition, since the cracked silt layers are detached from the underlying sand, they do not prevent water from reaching the sand layer and exposing it to infiltration. In contrast, the flood plains, sandy inland islands, and terraces are characterized by overbank deposits of alternating thin layers of sand and silt-clay that dramatically reduce infiltration through them when flooded. The observation that flood water infiltration is less sensitive to flood stage has also been reported from numerical simulation (Bailey 2002; Blasch 2004).

A common observation on flood hydrology of arid environments is that the large floods with high peak water levels produce high transmission losses (Enzel 1990; Schick 1988). Accordingly, it is common to equate high transmission losses with high ground water recharge values (Greenbaum et al. 2001). However, in the Kuiseb River, the observed high transmission losses during high peak levels do not contribute much to ground water recharge. This claim is based on the observation that large floods may exceed the main active stream channel's boundaries and cover flood terraces, islands, and flood plains. Thus, a larger surface area is exposed to infiltration. During the short interval of high peak discharges, large quantities of water infiltrate through the flood plains and terraces. However, several observations in this study suggest that this wide flooding does not generate large recharge:

a) As shown earlier, during the first flood, when water infiltrated through dry sediments, the wetting-front propagation velocities did not exceed a few centimeters per hour (Figure 7b; Table 2). Accordingly, a flood peak stage that lasts for only a few hours on the flood plains will result in only shallow depth of infiltration. This infiltrating water will not reach the water table that is several meters further down.
b) Alternations of clay and sand beds are even more frequent in floodplain stratigraphy. Therefore, the vertical hydraulic conductivity of the floodplain is much lower than the hydraulic conductivity of the sandy active channel.
c) Even if we ignore the low infiltration potential of the flood plains and assign them the same infiltration rates measured in the stream channel, the short flooding duration minimizes the potentially deep infiltration into the dry sediments.

Therefore, we propose that even if the flood loses large quantities of water to infiltration into the very dry sediments on the flood plains, in the Kuiseb River setting and probably other hyperarid channels, only insignificant recharge to the alluvial aquifer is gained by this process. As shown by our results, the duration of the flood is more important and, therefore, the importance of large floods in hyperarid channels lies in the positive relationships in a specific drainage basin between flood magnitude and duration. Floods in arid areas with smaller peaks are usually of shorter duration than floods with larger peaks (Schick 1988). This is borne out mainly because of the characteristic shape of hydrographs in arid areas; the higher the peak, the longer the recessional limb of the hydrograph. As a result, it is not the short-term flooding of the flood plains and terraces that is important for recharge, it is the longer duration of the entire flood in the active channel.

Summary and Conclusions

A study on flood water infiltration and ground water recharge was conducted in the hyperarid section of the Kuiseb River, western Namibia. The study site was selected to represent a typical desert ephemeral river. In this study, we have implemented a novel experimental setup that allowed complete monitoring of water infiltrating from land surface down through the entire vadose zone to the ground water. In 2006, a rare wet summer season in west-central Namibia resulted in five floods during the relatively short period of 2 months, allowing a detailed investigation of the dynamic processes governing flood water infiltration and ground water recharge of shallow alluvial aquifers.

The monitoring system, which included FTDR probes installed in multiple depths along vadose zone cross section, combined with water level measurements of the flood and the ground water, allowed the following observations and conclusions:

- Each flood running on the channel bed initiated an infiltration process which was expressed through a sequential increase in vadose zone water content and, ultimately, a rise in ground water level as a final indication of the recharge process (Figure 7). The wetting sequence and timing allowed direct calculation of the wetting-front propagation velocity and infiltration fluxes for each individual flood event (Table 2).
- Three independent methods were used to calculate the flux: (1) the wetting-front propagation velocity with the observed changes in water content; (2) the rate of water table rise; and (3) the total increase in ground water storage. The similarity in the results calculated by these three methods reinforces the reliability of the technique and the validity of the data and conclusions (Table 2).
- Although each flood was characterized by different flow conditions (such as the initial water content of the sediments comprising the vadose zone), flood stages, and durations, very similar infiltration fluxes were calculated for all floods. We propose that this similarity is regulated at the top of the vadose zone near the channel bed; it is suggested that the almost unnoticeable microlayering of the sandy alluvial sediments at the top of the vadose zone governs the infiltration process and regulates the fluxes. As a result, the potential effect of high flood stages during large floods is buffered (Table 3).
- Throughout all of the floods, the entire vadose zone remained unsaturated with relatively low moisture content (10% to 15%). Saturation was achieved only with the rising water table following each infiltration event (Figure 5). This observation supports the suggestion that infiltration is regulated by the sediments near the surface.
- Small floods are not less important than larger floods since similar fluxes were recorded for both. Accordingly, it is suggested that above a certain flood stage threshold (probably ~15 to 25 cm) that allows infiltration, it is the flow duration and not the water stage that controls the recharge amounts. The impact of large floods on increasing total recharge comes from the general observation in hyperarid areas that larger floods are usually characterized by longer flow duration. Therefore, larger floods extend infiltration time from the bottom of the active channel (Table 3).
- Floods in arid regions that occasionally cover the flood plains will result in a larger proportion of transmission losses from the flood and in lower flood discharge downstream. These flooding zones, beyond the width of the active channel, will contribute very little, if any, additional water to ground water recharge.
- It is the limits on ground water storage capacity combined with the regulated maximum flux rate through the riverbed that control the ground water recharge. Most of the infiltration will take place in the main active channel under regulated fluxes, independent of the flood's stage. The storage capacity is determined by the width of the aquifer and the thickness of the unsaturated zone. When the water table reaches the surface underneath the active channel, the entire vadose zone is saturated and no more recharge can take place.

Acknowledgments

The authors would like to express their appreciation to Harmuth Kolb, from the Desert Research Foundation of Namibia and Michael Kugel and Yuval Shani from the Ben Gurian Universit of the Negev, who built, installed, and maintained the monitoring system at the Gobabeb-Kuiseb River site. We also wish to thank Rebecca Smith, Daniel Larsen, and Radu Bughici, who improved this manuscript through their thorough review. The study was funded by the sixth framework of the European Commission (WADE project, GOCE-CT-2003-506680), and by the U.S.–Israel Cooperative Development Research Program, Bureau for Economic Growth, Agriculture, and Trade, U.S. Agency for International Development (Grant No. C24-26).

References

Bailey, M.A. 2002. Analysis of one dimensional vertical infiltration using heat as a tracer in rillito creek. M.Sc. diss., The University of Arizona.

Blasch, K., T.P.A. Ferre, J. Hoffmann, D. Pool, M. Bailey, and J. Cordova. 2004. Processes controlling recharge beneath ephemeral streams in southern arizona. In *Processes Controlling Recharge Beneath Ephemeral Streams in Southern Arizona*, ed. J.F. Hogan, F.M. Phillips, and B.R. Scanlon, 69–76. Washington, DC: American Geophysical Union.

Botes, A., J. Henderson, T. Nakale, K. Nantanga, K. Schachtschneider, and M. Seely. 2003. Ephemeral rivers and their development: Testing an approach to basin management committees on the Kuiseb river, Namibia. *Physics and Chemistry of the Earth* 28, 853–858.

Bueno, A., and D.J. Lang. 1980. Aquifer discharge from the 1969 and 1978 floods in the Mojave river basin, California. USGS Water Resource Investigation Open File Report. Denver, Colorado: USGS.

Dahan, O., E.V. McDonald, and M.H. Young. 2003. Flexible time domain reflectometry probe for deep vadose zone monitoring. *Vadose Zone Journal* 2, no. 2: 270–275.

Dahan, O., Y. Rimon, B. Tatarsky, and T. Ruti. 2006. Deep vadose zone monitoring system. In *TDR 2006: 3rd International Symposium on Time Domain Reflectometry for Innovative Soils Applications, Purdue University, USA.*

Dahan, O., Y. Shani, Y. Enzel, Y. Yechieli, and A. Yakirevich. 2007. Direct measurements of floodwater infiltration into shallow alluvial aquifers. *Journal of Hydrology* 344, no. 3–4: 157–170.

de Vries, J., and I. Simmers. 2002. Groundwater recharge: An overview of processes and challenges. *Hydrogeology Journal* 10, no. 1: 5–17.

Enzel, Y. 1990. Hydrology of a large, closed arid watershed as a basis for paleohydrological and paleoclimatological studies in the Mojave river drainage system, southern California. Ph.D. diss., University of New Mexico.

Gee, G.W., and D. Hillel. 1988. Groundwater recharge in arid regions: Review and critique of estimation methods. *Hydrological Processes* 2, 255–266.

Greenbaum, N., U. Schwartz, A.P. Schick, and Y. Enzel. 2001. Paleofloods and the estimation of long-term transmission losses and recharge to the lower Nahal Zin alluvial aquifer, Negev desert, Israel. In *Paleofloods and the Estimation of Long-Term Transmission Losses and Recharge to the Lower Nahal Zin Alluvial Aquifer, Negev Desert, Israel*, ed. P.K. House, D. Lavish, R.H. Webb, and V.R. Baker, 311–328. Washington, DC: American Geophysical Union.

Jacobson, P.J., K.M. Jacobson, and M.K. Seely. 1995. Ephemeral rivers and their catchments: Sustaining people and development in western Namibia. Windhoek, Namibia: Desert Research Foundation of Namibia.

Lange, J. 2005. Dynamics of transmission losses in a large arid stream channel. *Journal of Hydrology* 306, no. 1–4: 112–126.

Osterkamp, W.R., L.J. Lane, and C.S. Savared. 1994. Recharge estimates using a geomorphic/distributed-parameters simulation approach, Amargosa river basin. *Water Resources Research* 30, no. 3: 493–507.

Rimon, Y., O. Dahan, R. Nativ, and S. Geyer. 2007. Water percolation through the deep vadose zone and groundwater recharge: Preliminary results based on a new vadose zone monitoring system. *Water Resources Research* 43, W05402, doi: 10.1029/2006WR004855.

Sanford, W. 2002. Recharge and groundwater models: An overview. *Hydrogeology Journal* 10, no. 1: 110–120.

Scanlon, B.R. 2004. Evaluation of methods of estimating recharge in semiarid and arid regions in the southwestern US. In *Groundwater Recharge in a Desert Environment: The Southwestern United States*, ed. J.F. Hogan, F.M. Phillips, and B.R. Scanlon, 235–354. Washington, DC: AGU, Water science and application 9.

Schick, A.P. 1988. Hydrologic aspects of floods in extreme arid environments. In *Flood geomorphology*, ed. V.R. Baker, R.C. Kochel, and P.C. Patton, 189–203. New York: John Wiley and Sons.

Scholz, H. 1972. The soils of the central Namib Desert with special consideration of the soils in the vicinity of Gobabeb. *Madoqua ser.* 2, 33–51.

Schwartz, U. 2001. Surface and near-surface responses to floods in a large channel (Nahal Zin) in the context of an alluvial aquifer in a hyper-arid environment. Ph.D. diss., Department of Geography, Hebrew University of Jerusalem.

Shentsis, I., and E. Rosenthal. 2003. Recharge of aquifers by flood events in an arid region. *Hydrological Processes* 17, no. 4: 695–712.

Shentsis, I., L. Meirovich, A. Ben-Zvi, and E. Rosenthal. 1999. Assessment of transmission losses and groundwater recharge from runoff events in a Wadi under shortage of data on lateral inflow, Negev, Israel. *Hydrological Processes* 13, no. 11: 1649–1663.

Simmers, I. 1997. *Recharge of Phreatic Aquifers in (Semi) Arid Areas*. Rotterdam, The Netherlands: AA Balkema.

Walter, M.T., J.S. Kim, T.S. Steenhuis, J.-Y. Parlange, A. Heilig, R.D. Braddock, J.S. Selker, and J. Boll. 2000. Funneled flow mechanisms in a sloping layered soil: Laboratory investigation. *Water Resources Research* 36, no. 4: 841–849.