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Review Paper

Transmission losses, infiltration and groundwater recharge through ephemeral and intermittent streambeds: A review of applied methods

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SUMMARY

Aquifer recharge through ephemeral streambeds is believed to be a major source of groundwater recharge in arid areas; however, comparatively few studies quantify this streamflow recharge. This review synthesizes the available field-based aquifer recharge literature from arid regions around the world. Seven methods for quantifying ephemeral and intermittent stream infiltration and aquifer recharge are reviewed; controlled infiltration experiments, monitoring changes in water content, heat as a tracer of infiltration, reach length water balances, floodwave front tracking, groundwater mounding, and groundwater dating. The pertinent temporal and spatial scales, as well as the advantages and limitations of each method are illustrated with examples from the literature. Comparisons between the methods are used to highlight appropriate uses of each field method, with emphasis on the advantages of using multiple methods within a study in order to avoid the potential drawbacks inherent in any single method. Research needs are identified, including: quantitative uncertainty analysis, long-term data collection and analysis, understanding of the role of riparian vegetation, and reconciliation of transmission losses and infiltration estimates with actual aquifer recharge.

Contents

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1. Introduction

As the global population increases, humans are increasingly putting pressure on the resources of arid regions, where water scarcity is a major issue (Seely et al., 2003). In these regions, aquifers are often the principal water supply, because surface water is unreliable and usually only present during or after flood events; however, the development of groundwater resources often proceeds without a thorough understanding of the recharge processes (Edmunds, 1998). It is frequently asserted that infiltration through streambeds during flood events is the main form of recharge (Sorman and Abdulrazzak, 1993; Abdulrazzak, 1995; Shentsis and Rosenthal, 2003; Subyani, 2004; Niswonger et al., 2005), though there have only been a relatively small number of studies that have quantified this rate. In many areas, groundwater resources may be fossil resources, and rates of extraction may greatly exceed rates of recharge. Because precipitation varies both spatially and on temporal scales of years to decades, extraction rates need to take into account long-term aquifer recharge, which is only complicated further by the sporadic nature of the recharge events themselves (Besbes et al., 1978). Although the actual amount of streamflow that reaches the underlying aquifer (i.e. recharge) is typically the quantity of interest, several factors make it difficult to measure this recharge. These factors include the often large depth to groundwater (potentially causing a long delay from infiltration to recharge and complicating efforts to collect data), spatial variability in recharge due to geologic heterogeneity beneath streams, and potential difficulties in linking a specific streamflow event to a change in aquifer level. Therefore, it is common to use methods that estimate transmission loss or infiltration as a proxy for measuring groundwater recharge. Transmission loss quantifies streamflow reductions including infiltration through the sediments, evapotranspiration back to the atmosphere, and loss to stream banks or floodplains as the water travels downstream. Infiltration rates are typically measured at or just below the streambed surface, which is easier to access than a deep water table. However, several other factors complicate infiltration measurements, specifically in ephemeral and intermittent streams. For example, precipitation is spatially variable, making it difficult to arrive on-site in time to collect data in typically remote areas, frequent scouring and deposition of the streambed during flood events makes it difficult to estimate streambed geometry, and flood events themselves are unpredictable and transient in nature (Pilgrim et al., 1988; Shannon et al., 2002). Partly because of these challenges, there is a wide body of literature characterizing infiltration through unsaturated soils or diffuse recharge, but fewer studies describe field results from the ephemeral streambeds themselves.

Review papers exist for groundwater recharge in general (De Vries and Simmers, 2002; Scanlon et al., 2002), but not specifically for the ephemeral and intermittent streams characteristic of arid systems. In this paper, we therefore synthesize the field-based studies that quantify transmission losses, infiltration, or aquifer recharge from ephemeral and intermittent stream systems. Much of the available literature describes the arid regions of western USA, with the findings of that research summarized in Hogan et al. (2004) and Stonestrom et al. (2007). Some of the arid systems of southwest Africa have also been studied extensively (Lange, 2005; Bauer et al., 2006; Dahan et al., 2008; Morin et al., 2009). A further area of concentrated research is in the wadis of Saudi Arabia (Abdulrazzak et al., 1989; Sorman et al., 1997; El-Hames and Richards, 1998; Wheater and Al-Weshah, 2002). In this summary, we aim to review the pertinent methods using examples from all of these areas. Although many more studies have examined the processes involved in ephemeral stream recharge using theoretical or laboratory studies, we focus here on field-based studies. Further, although diffuse recharge throughout the catchment can also play a role in aquifer recharge during precipitation in arid zone catchments, for simplicity we focus on the methods and examples related specifically to streambed infiltration or recharge. Each of the appropriate methods is presented in terms of what is being measured (i.e. infiltration, transmission loss, or recharge), spatial and temporal ranges, and advantages and limitations. To illustrate cases where each of these techniques is relevant, we review the pertinent studies describing field applications of each method. The methods are then compared, and considerations common to all of the methods are illustrated with further examples. Finally, potential research gaps and future directions are suggested.

2. Methods

Several methods have been used for quantifying loss rates from losing streams, and some of these have recently been reviewed by Kalbus et al. (2006). However, not all of the methods that can be used in perennial streams are appropriate for ephemeral or intermittent streams. The techniques that are generally available to study ephemeral systems can be divided into three groups. The first group of methods monitors infiltration through the streambed. These methods typically provide point estimates of infiltration. They include:

(1) Controlled infiltration experiments.
(2) Monitoring changes in water content.
(3) Heat as a tracer of infiltration.

The second group of methods is based on measurements of streamflow during flow events. These methods provide estimates of either transmission losses or streambed infiltration over much larger spatial scales, sometimes up to several tens of kilometers of river distance. The methods include:

(4) Reach length water balance.
(5) Floodwave front tracking.

The third group of methods is based on measurements within the groundwater underlying the ephemeral stream. These methods therefore provide estimates of actual groundwater recharge, rather than streambed infiltration. These estimates will usually represent spatial and temporal averages. The methods include:

(6) Groundwater mounding.
(7) Groundwater dating.

The broad principles of each method are described in the following sections, together with their advantages and challenges. The reader is referred to the cited literature for detailed technical descriptions of each method, as these are described elsewhere and because the equations and methods can vary widely even within one category.

2.1. Controlled infiltration experiments

Controlled experiments in dry channels can be used to estimate infiltration rates during flood events. These experiments typically involve creating a column of constant head above the streambed and directly measuring the rate of infiltration, from which soil properties such as sorptivity and field saturated hydraulic conductivity can be calculated. This can be achieved using an infiltrometer or permeameter for measurement at a certain location within the streambed, or by isolating and filling a relatively short reach of the stream (typically less than 50 m). Infiltrometers are typically metal cylinders that are placed on the streambed, in which a constant water level can be maintained by addition of water. Similarly, a permeameter is a cylinder which can be used to keep a constant level of water in contact with the soil (constant head) at a specific depth.

Dunkerley (2008) used a ring infiltrometer to measure infiltration into the bed and banks of an ephemeral stream in New South Wales, Australia. Water levels were monitored inside the tilted cylinder, which was inserted each 0.5–1.0 m up the bank in three vertical transects. He found that infiltration rates increased with distance up the stream bank (ranging from 0.02 to 1.8 m/d vertical distance above the streambed), especially where the presence of a mud drape (fine sediments deposited by previous flows) was absent. As an example of a controlled ponding experiment, Batlle-Aguilar and Cook (2012) artificially flooded a 7 m length of intermittent stream in southern Australia to calculate stream infiltration rates and to determine how infiltration rate varied with stream depth. Infiltration rates were found to vary between 0.3 and 1.8 m^2/d for water depths between 0.2 and 0.38 m.[1] The authors estimated that flow events of 10–15 days were required to generate aquifer recharge, while the infiltration from shorter flow events would be completely lost to evapotranspiration.

While these dry channel experiments are not suitable for measuring infiltration from flood events directly, they provide a means for estimating streambed hydraulic properties in controlled experiments where it is not convenient to sample during natural wetting events. Further, it can be impractical to get head measurements in variably saturated sediments beneath ephemeral streams (Ronan et al., 1998), whereas infiltrometer or permeameter measurements can be repeated many times within a reach to provide an estimate of the variability of infiltration rates. However, several studies suggest that the hydraulic properties or infiltration rates measured during controlled infiltration experiments may not always be indicative of reach-scale infiltration rates during flood events. Crerar et al. (1988) found that silt carried by floodwaters can effectively seal the surface of the streambed, greatly reducing the infiltration rate. This process occurred even at relatively high flow velocities, where sand grains were mobilised. Thus, the hydraulic conductivity measured between flow events is only a snapshot estimate and does not capture the often transient nature of infiltration rates during ephemeral stream flow events. In comparing estimates using a ring infiltrometer during a controlled experiment with transmission losses during a natural flood event, Dahan et al. (2007) observed that infiltration rates during the flood were only half of those measured with the infiltrometer. The authors found that water traveled quickly through preferential pathways during the natural flood event, although the quantity transferred through matrix flow was greater in the infiltrometer experiment. They also observed that the sediments beneath the streambed but above the water table never reached saturation; therefore, use of saturated hydraulic conductivity to calculate infiltration would overestimate actual values. Further, Dunkerley (2008) showed that direct measurements may under or over-estimate expected infiltration rates during flood events, because they cannot capture the spatial variability associated with scouring and depositional processes, which may greatly affect infiltration.

2.2. Monitoring changes in water content

This method estimates the infiltration rate from the advance of a wetting front as it travels vertically through streambed sediments in response to a wetting event (streamflow). Information on the change in stored water in the vadose zone (i.e. changes in moisture content) and the velocity of the wetting front through the soil column allows calculation of water flux through the sediments. Therefore, this method is exclusive to ephemeral or intermittent streams, where the streambed is initially unsaturated or partially saturated.

Although there is a wide body of literature describing infiltration through unsaturated soils and the measurement of water content has been cited as the common method for determining ephemeral stream recharge (Dowman et al., 2003), actual examples of this method used in natural ephemeral streams are rare, partially due to the need for access tubes (Constantz and Thomas, 1997). For example, Parissopoulos and Wheater (1992) used a neutron probe in a 3 m deep access tube to record water content versus depth both before and during a controlled infiltration experiment. By matching the wetting curves, they were able to determine soil hydraulic properties within several layers of soil in a wadi in Saudi Arabia. Although they did not estimate infiltration rates, they were able to predict high infiltration rates based on the observed high hydraulic conductivity of the soil profile.

Water content measurements allow quick detection of the vertical rate of infiltration in response to a flood event. Once the streambed sediments at the measurement depth are saturated, changes in flux cannot be determined from water content. Thus, a combination of methods is required to determine both short and long-time infiltration flux in medium to long flood events where late time infiltration can be a significant source of groundwater recharge. As observed in Dahan et al. (2008), transmission losses are lower as the streambed sediments approach saturation (or a maximum saturation degree, as streambed sediments often do not reach saturation until the water table rises), making it especially important to capture the short-time infiltration in shorter duration flood events.

Measurements of water content are typically collected at one point along the length of a stream reach. However, unlike temperature sensors (see below), water content sensors typically cannot be deployed without the use of a separate data logger, making it typically expensive to deploy sensors at many locations along the longitudinal extent of a river. A wide variety of geophysical methods is now available to determine vertical water contents and changes in water content over time (Ferré et al., 2007). Increased use of these methods has potential to expand our understanding of infiltration processes following flood events, although care must be taken where streambed salinity is also spatially and temporally variable.

Dahan et al. (2007, 2008) developed a monitoring system of time domain reflectometry probes attached to flexible tubing that helps alleviate the problems associated with installation of the sensors. Using this vadose zone monitoring system in conjunction with piezometers and water level sensors to capture changes in groundwater and surface water levels, respectively, Dahan et al. (2008) were able to independently calculate ephemeral recharge using the wetting front propagation in the vadose zone, the rate of rise of the water table, and the change in groundwater storage. Over five flood events of varying duration ($\sim$1 day to almost 2 weeks) and magnitude, the authors measured infiltration fluxes of 0.17–0.36 m/d, corresponding to changes in groundwater storage of 0.17 – 0.22 m/d.

2.3. Heat as a Tracer of Water Movement

Although more commonly used in perennial streams to determine flux rates through the streambed, several studies have used measurements of temperature at the surface and at various depths within the streambed to determine recharge through ephemeral channels. This method relies on the transmission of a transient temperature signal propagating from the surface of the streambed

[1] Infiltration estimates are a measure of water flux, which is a volume per area per time. This is commonly simplified to length per time; however, where infiltration has been measured as a volume in a particular stream reach, it may also be expressed as volume per length of channel per time (as in the m^2/d here) or simply total volume per time (such as m^3/d).

into the sediments. When there is no water flowing in the ephemeral or intermittent stream, this signal is propagated only through conduction. During periods of streamflow, both conductive and convective processes are present, allowing percolation characteristics beneath the streambed to be evaluated (Constantz and Thomas, 1997) and infiltration rates to be calculated. Periods of streamflow are usually easily discernible from abrupt changes in the shape of the stream temperature record (Constantz et al., 2001) (Fig. 1).

Much of the work to date using temperature as a tracer of ephemeral stream recharge has focused not on quantifying the actual recharge, but on identifying the length and/or depth of percolation within the river. In a pioneering study, Constantz and Thomas (1997) investigated diurnal and seasonal changes in temperatures at depths between 0.3 and 3.0 m beneath two arroyos in central New Mexico, USA. While one of the arroyos clearly displayed convective heat transfer past the extent of the root zone (i.e. 3 m) and therefore contributing to groundwater recharge, the other arroyo only wetted part way through the profile in response to monitored flood events. Blasch et al. (2004) developed a method of analyzing the moving standard deviation to estimate wetted periods just below the surface and at greater depths and used the method to identify streamflow in an Arizona creek that typically flows for less than 24 h at a time. Brown et al. (2006) used long-term temperature data from in and beneath mountain streams in the French Pyrenees to identify periods of dewatering and/or freezing and therefore determine which stream reaches showed annual flow permanence. In several ephemeral streams in southwestern USA, changes in temperature were used to identify gaining, losing, and dry conditions; combined with streamflow records and vertical profiles of environmental tracers, the authors were then able to identify periods of infiltration and estimate overall annual recharge (Moore, 2007; Stewart-Deaker et al., 2007; Stonestorm et al., 2007).

In an early use of temperature measurements to quantify the infiltration beneath an ephemeral channel, Ronan et al. (1998) installed thermocouples down to a depth of almost 3 m beneath the right and left banks and beneath the center of an ephemeral stream in western Nevada, USA. They fit a two-dimensional, variably-saturated, numerical water and heat flow model to streambed temperature data from two periods of streamflow to calculate infiltration rates on the order of 1.2 m/d beneath the streambed, which agreed well with estimates from differential streamflow gauging. Temperature traces collected below the stream bank could not be adequately fitted using the two-dimensional representation, and the authors concluded that three-dimensional thermal patterns

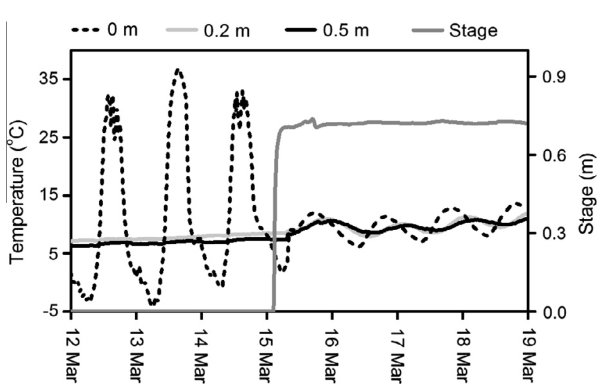

Fig. 1. Temperature data from a large channel in eastern Colorado, USA (Shanafield, M, unpublished data), showing patterns of streambed temperature at depth below the surface both before and during a flow event. The data at zero meters shows air temperature until water filled the channel on 15 March, and surface water temperatures thereafter. Once infiltration begins, advection of water results in a daily sinusoidal pattern in the temperatures at depth.

due to down canyon water flow through the unsaturated zone were responsible for the observed patterns in temperature. As an example of the use of this method over longer timescales, Kulongoski and Izbicki (2008) used access tubes beneath two washes in the Mojave Desert, USA, to examine annual variations in temperatures down to 30.5 m below ground. Using this data in a numerical heat transport model, they were able to estimate average annual infiltration rates of 2–4 m/y, with the frequency of infiltration occurring only every 3 years on average.

The ability to bury temperature sensors within the streambed makes them useful in environments where flood events cause significant scouring and can damage equipment installed at the streambed surface. Another advantage of using temperature measurements to capture infiltration rates is the temporal versatility of the sensors. Sensors can be attached to data loggers that commonly sample temperatures on the order of seconds and average the data into 5–30 min measurements (Ronan et al., 1998; Constantz et al., 2001, 2002; Blasch et al., 2004). Using a constant infiltration experiment, Dowman et al. (2003) found that for depths greater than 1 m, temperature measurements collected in a cased borehole before and after an infiltration event matched temperatures collected from thermocouples buried directly in the porous medium and could therefore be used to estimate recharge during a flood event. However, several hours were required for the monitored borehole temperatures to equilibrate, making this method better suited to longer flow events. In a study of several ephemeral streams and arid basins, Constantz et al. (2003) used borehole temperatures down to a depth of almost 250 m, showing the ability of this method to capture long-term (period of years) recharge patterns. However, the measured recharge at these depths is typically indicative of overall, diffuse recharge, not focused recharge from streamflow events.

Care should be taken to assure that temperature changes during flood events can be attributed to advective water movement. In a study of flood events along the Kuiseb River, Amiaz et al. (2011) observed thermal anomalies in the unsaturated zone caused by pressure pulses generated by the flood at the surface. These anomalies were attributed not to heat transfer by flowing water, as is generally assumed when using temperature measurements to estimate infiltration, but to instant variations in the pressure of the gas phase in the unsaturated zone during the flood. The multi-dimensional nature of subsurface flow also remains a challenge in using heat as a tracer to quantify the infiltration beneath ephemeral streams. Shan and Bodvarsson (2004) developed a layered, one-dimensional analytical solution that can help to estimate the non-vertical, steady, lateral flow of water by calculating differences in infiltration beneath vadose zone layers with differing thermal properties (i.e. below the temperature and water content extinction depth). However, this requires good knowledge of the sediment thermal properties, which is often missing from field studies. Using vertical temperature profiles in several ephemeral rivers in southwestern USA, Constantz et al. (2002) calculated initial and steady state vertical infiltration rates of approximately 0.1–0.4 m/d and 0.75–2.0 m/d, respectively. The authors estimated that the channel loss calculated by scaling up one-dimensional, temperature profile estimates of percolation were 30–50% less than loss rates calculated from streamflow gauging. This difference was attributed to lateral flow through the streambed, which is not captured in the one-dimensional temperature tracer method, and error in estimating wetted stream channel area for scaling up the point estimates obtained at specific profiles. Because temperature sensors are relatively inexpensive, the error caused by collecting data at a single profile can be minimized by putting in many profiles, which can be spread out over a relatively long longitudinal reach of a stream. For example, Constantz et al. (2001) mapped streamflow in 10 ephemeral streams in southwest USA by placing temperature sensors up to tens of kilometers apart. With these, the authors could determine how far down each stream the flood events succeeded in producing flow. Alternatively, many temperature probes can be placed within a small cross-section of streambed, allowing multi-dimensional patterns of infiltration to be discerned (Ronan et al., 1998).

2.4. Reach length water balance

Streamflow differencing has been widely used for estimating transmission losses in perennial streams. When flow is stable, loss rates are relatively easily accessed by measuring the difference between upstream and downstream flow (Harte and Kiah, 2009; Schmadel et al., 2010), while taking into account other flow sources and sinks, including evaporation. In ephemeral streams, flow is rarely stable, and so instead loss rates are determined by integrating the flow rate from the upstream and downstream stations across the entire flow event (Abdulrazzak, 1995). Flow is typically measured automatically, using rating curves to relate flow depth to discharge rate, but can also be measured using acoustic Doppler current profiling (ADCP) or manual techniques. More recently, satellite imagery has been used to discern flow loss rates along a stream reach (Walter et al., 2012). Where a sufficiently long period of record is available and a number of flow events are captured, relationships between transmission loss and flow volume can be derived (Lane et al., 1971; Walters, 1990).

Determining accurate rating curves for gauging stations in ephemeral streams is challenging. Streamflow in ephemeral and intermittent stream basins is often quite flashy in nature, leading to significant scouring and deposition in the streambeds and causing changes in streambed geometry and subsequent inaccuracies in gauging stations and rating curves (Constantz and Thomas, 1997). Further, fieldsites are often remote from population centers, and the unpredictable nature of flood events therefore makes it difficult to take manual readings for calibrating the rating curves regularly. Because of the relatively high error therefore associated with streamflow measurements, upstream and downstream gaugings must be spaced far enough apart so that the error is not greater than the transmission loss. Reaches on the order of 4–30 km are therefore not uncommon, leading to overall transmission loss estimates of up to 100,000 m^3 of water during a single flood event only 2–5 h in duration (Lane et al., 1971; Walters, 1990). In many cases, calculation of transmission losses in these long reaches is difficult, because tributaries enter the main channel between gauging stations, and these are often poorly gauged. In some cases, sporadic gaugings in the tributaries can be used to establish relationships between tributary inflow and flow in the main channel (Osterkamp et al., 1995). Alternatively, relationships between transmission losses and other hydraulic parameters (e.g., channel width) can be established, and this can allow extrapolation of results of ungauged channel reaches (Walters, 1990). However, the relatively high spatial variability of rainfall, which often complicates the prediction of streamflow timing in these arid systems (Shannon et al., 2002; Mudd, 2006), can make establishing these relationships difficult (Shentsis et al., 1999).

Another challenge with this method is the estimation of evaporation, including soil evaporation. This can be particularly difficult if the river flow leaves the channel and spreads out across a floodplain. Across a large area, relatively small errors in evaporative loss can lead to large errors in estimated infiltration.

In an innovative use of differential gauging, Walter et al. (2012) used digital orthophoto quadrangles (DOQs) and Systeme Probatoire de l'Observation de la Terre (SPOT) images to estimate channel width, performed surveys of the channel at several cross-sections to estimate depth, and computed velocity using a power function. These inputs were used in Manning's equation to solve for flow at several ungauged stations downstream of the one gauged station. From these estimates they were able to determine which areas of the stream reach had highest loss rates. In another use of satellite imagery, Costa et al. (2012) combined gauged flow measurements with satellite imagery to determine streambed geometry and river dynamics between stream gauges to estimate relationships between input flow and transmission losses for 60 km of the Middle Jaguaribe River, Brazil. Using this method they were able to develop an empirical relationship for transmission losses over a range of input flows. Calculation of transmission loss from differential gauging also allowed Pool (2005) to evaluate long-term patterns of recharge in three streams in Arizona, USA. These transmission losses were combined with seasonal precipitation and groundwater storage change data to understand the influence of El Nino Southern Oscillation climatic patterns on regional recharge and to estimate long-term changes in infiltration.

2.5. Floodwave front tracking

Tracking the advancement of a floodwave down an ephemeral or intermittent stream channel allows transmission loss or infiltration to be estimated based on the velocity, and changes in velocity, along the longitudinal direction of flow (Fig. 2). For example, if none of the water traveling downstream during a flood event could infiltrate, the velocity of water traveling down the channel would be constant (assuming constant geometry). Infiltration losses represent a mass sink that alters this velocity, and can therefore be estimated, assuming inflow and channel geometry are known. The equations used for modeling floodwave progression in ephemeral streams vary widely, from the simpler Muskingum–Cunge method (Cunge, 1969) to the full Saint–Venant equations, which combine the momentum and continuity equations (Chow et al., 1988). The simpler equations, such as the kinematic wave, typically have the advantage of requiring less input data, although they may not accurately represent the system when pressure or acceleration forces cannot be ignored, such as in mild-sloped rivers (Chow et al., 1988). These models are typically calibrated to the conditions as the stream wets, but then also used to estimate recharge under long-time flow conditions or as drying occurs. This makes them suitable for long duration flow events.

Floodwave routing models of various complexities have been used on ephemeral rivers across the world. Lange (2005) used the Muskingum–Cunge method to determine the relative importance of channel transmission losses over 150 km of the ephemeral Kuiseb River in Namibia. Differencing of streamflow gauging stations along the river indicated transmission losses of between 0.8 and 22.5×10^6 m^3 per flood event, equivalent to between 5 and 150 m^3/m. They found that large flood events incurred higher

Fig. 2. A wetting front moving down an initially dry channel in eastern Colorado, USA. The speed of the advancing front is sensitive to the hydraulic conductivity of the streambed (and hence the infiltration rate), allowing for this to be calibrated based on observed timings of front arrival along the channel.

transmission losses than small to medium floods, potentially due to overbank flooding or the disturbance of a surficial clogging layer at high discharges. El-Hames and Richards (1998) coupled the Richards equation for infiltration with kinematic wave routing and used the model to determine an average transmission loss of 16×10^6 m^3/m in 4–7 h over 23 km of Wadi Tabalah in Saudi Arabia. Due to the inability of the kinematic wave to match floodwave timings in several large, very mild sloped channels in Colorado, USA, Shanafield et al. (2012) used the more complicated diffusion wave analogy to better capture momentum losses along 0.1–0.4 km segments of the channel. In an application of this model, Noorduijn et al. (2014) were able to match both the floodwave advancement along the surface as well as the groundwater response for a large open channel in southeastern Australia. The authors estimated long-term seepage loss rates varying between 10^{-3} and 10^{-7} $m^3\,d^{-1}\,m^{-2}$ over a 1.4 km reach, and showed that the use of groundwater information, in addition to the timing of the flood wave advance along the surface of the channel, allowed for better characterization of the seepage flux into the channel bed.

An advantage of the floodwave front tracking method is its ability to capture a wide spatial range. For example, Niswonger et al. (2008) estimated hydraulic conductivities and identified areas of high transmission loss over 43 and 11 km of an intermittent and an ephemeral stream channel, respectively. Similarly, Morin et al. (2009) estimated recharge over stream reaches up to 55 km long using a kinematic wave model. Alternatively, Noorduijn et al. (2014) used a floodwave model to characterize spatial differences in infiltration rates for each 100 m over 2 km of an ephemeral channel.

One challenge with this method is in the potential sensitivity of the models to the stream geometry, which can be difficult to adequately parameterize over long reaches and can be altered during flood events. Noorduijn et al. (2014) found that for seepage rates below approximately 10^{-4} $m^3/d/m^2$, uncertainty in the Manning's roughness parameter and the wetted width became important, demonstrating the need for an accurate understanding of stream geometry when using this method. Further illustrating this point, Costa et al. (2012) devised a multi-step model using the continuity equation to route water down the channel with a term for potential infiltration, which was subsequently partitioned into lateral groundwater flow, unsaturated seepage, and soil water redistribution. The model was applied to a 1.5 km section of stream in the Walnut Gulch Experimental Watershed to predict transmission losses, and was found to consistently underpredict flood event volumes, possibly due to limited knowledge about stream geometry.

2.6. Groundwater mounding

The formation of a groundwater mound occurs when water infiltrating through a streambed reaches the aquifer. The magnitude of the mound depends on the ratio of the recharge rate to the rate that the aquifer's transmissivity allows the water to move away laterally. The changes in groundwater level close to the stream can be monitored relatively easily using monitoring wells. The response of this groundwater mound to changes in streamflow can be used to calculate changes in the volume of groundwater storage, and therefore infiltrated stream water that has recharged the aquifer. Widely used analytical solutions include Hantush (1967) and Bouwer (1969). This approach makes the assumptions that percolation beneath the stream is downwards to the water table, and that the aquifer is homogeneous and isotropic (Sorman et al., 1997). Moench and Kisiel (1970) developed perhaps the first convolution relation for examining aquifer recharge. It uses the height of a groundwater mound above a static water table due to an instantaneous slug of recharge as the input function and the observed time series of water levels in a well located a distance from the channel as the output function for estimating recharge to the aquifer per unit channel length.

There are many examples of the use of groundwater mounding equations in the literature, yielding a wide range of recharge estimates. Besbes et al. (1978) estimated aquifer recharge for a large flood event in the Kairouan basin of Tunisia. For large floods (volume of at least 2,000,000 m^3) the authors were able to use the available monthly groundwater head data to successfully estimate aquifer recharge using a convolution approach. They estimated recharge volumes of 6×10^7–12×10^7 m^3 beneath two wadis over a four month period. Sorman et al. (1997) compared a numerical model of the one-dimensional Boussinesq equation to an analytical approach to estimate recharge from three storm events, each of approximately one day in duration, in a wadi in Saudi Arabia. They found that the techniques were comparable for both steady-state and transient recharge rates beneath the wadi for the observed recharge rates of 0.48–1.44 m/d. Using similar methods, Stephens et al. (1998) compared steady-state infiltration rates into two stream channels in New Mexico, USA with the Bouwer (1969) analytical and the Moench and Kisiel (1970) convolution methods. Using the Bouwer (1969) method and assuming no clogging layer yielded an infiltration rate of 5.5 m/d, four times larger than the estimate of 1.4 m/d without a clogging layer., and resulting in almost four times as much total infiltration (12.0 versus 3.0 m^3/m). The convolution method yielded estimates that were within this range, with an infiltration rate of 5.0 m/d and a total recharge estimate of 10.7 m^3/m over the 40 week study period. Fulton (2012) used the Hantush (1967) analytical solution to estimate the recharge rate from the Finke River, central Australia, based on the transient groundwater mound which developed following an October 2010 flood event. The water table in a bore on the edge of the river was 65 m beneath the stream before the flow event, and rose 3.4 m following the flow event, giving a total recharge of 1.3 m over the 8.5 day event.

An advantage of this method for estimating ephemeral recharge is that because changes in the groundwater levels themselves are used, the method yields actual recharge estimates and not infiltration, which can be considerably greater. Further, the method can be used for flood events of relatively long duration (e.g. 40 weeks in Stephens et al., 1998). The disadvantages of this method concern the many assumptions that must be made. Consideration of heterogeneity within the porous medium and changes in geological properties in the longitudinal direction require detailed measurements of groundwater head and sediment properties for input into a numerical water transport model. Uncertainties in hydraulic conductivity or specific yield are likely to lead to uncertainties in estimated fluxes of at least ±50%. For example, Goodrich et al. (2004) used the Hantush (1967) equation to reproduce observed water levels at two wells in southwestern USA. They found that a unique combination of initial recharge rate and hydraulic conductivity was necessary at each well, despite the assumption of an isotropic and homogeneous aquifer. For the 2-year study period, their groundwater mounding model estimated roughly half the volume of total recharge predicted by using the water balance, heat tracer methods, and roughly one-third less recharge than micro-gravity measurements. Another common assumption for this method is that of Dupuit flow, which can lead to errors in groundwater level calculations when the observation well is very close to, or beneath the stream (Brunner et al., 2009). Also, the method relies on changes in recharge to generate a transient groundwater mound. Where the unsaturated zone is deep, temporal variations in infiltration are damped during passage through the unsaturated zone, so that the temporal variations in recharge are less than the variations in infiltration. This may result in the method underestimating the recharge rate.

New techniques for measuring the change in groundwater storage, such as microgravity, are now in use. Gravity measurements were used by Pool and Schmidt (1997) to estimate a total recharge of 13.4×10^6 m^3 near Rillito Creek, Arizona, USA, for a period of just over one year, approximately 90% of which resulted from a series of winter flood events. In a further study of Rillito Creek, Hoffmann et al. (2007) used gravity measurements to estimate changes in storage of 7.5×10^6 and 11.1×10^6 m^3 for a 14 km reach, which compared well with recharge estimates using vertical infiltration from Darcy's Law calculations, a water balance, and one- and two-dimensional models using temperature as a tracer. Gravity measurements were also used to characterize riverbank infiltration conductances for an ephemeral river in the Okavango Delta in Botswana in response to flood events (Christiansen et al., 2011).

Gravity is the only available method that directly senses mass change (which is normally attributed to change in water storage), and has the advantage of being non-intrusive and not influenced by small-scale heterogeneity. The disadvantages are that ground-based microgravity measurements must still be scaled up from the cross-section and require complicated data interpretation due to the non-uniqueness of the inverse problem (Goodrich et al., 2004), while satellite-derived data is largely still of a scale too coarse for localized measurements (Goodrich et al., 2004; Scanlon, 2004; Walter et al., 2012).

2.7. Groundwater dating

The groundwater velocity away from a losing stream in response to flood events can be determined from measurements of groundwater age derived from environmental tracers. While other chemical methods or tracers, including salinity or stable isotopes, may be useful for distinguishing between sources of water in the aquifer, they can only be used to determine actual flux rates if individual precipitation events could be identified by a unique chemical composition. A range of environmental tracers can be used to estimate groundwater age (or residence time) over timescales ranging from days to thousands of years (Cook and Böhlke, 2000). The rate of increase in groundwater age with distance from the river allows calculation of the groundwater velocity. Many different tracer methods have been used for this purpose, including radon (Hoehn and Von Gunten, 1989; Bertin and Bourg, 1994), tritium (Geyh et al., 1995), ^{3}H/^{3}He (Stute et al., 1997; Massmann et al., 2009), and ^{14}C (Drury et al., 1984; Fulton, 2012). Although these techniques have been more commonly used to measure infiltration rates from perennial rivers, they are equally applicable to ephemeral rivers.

The choice of tracer will depend upon the likely rate of recharge and the distance of the available bores from the river. Hoehn and Von Gunten (1989) used radon activities in groundwater to estimate subsurface residence times between 2 and 8 d in bores between 2 and 30 m from the River Glatt, Switzerland. Groundwater residence time increased with distance from the river, consistent with the river being the source of the local groundwater. Based on the rate of increase in residence time with distance, the authors estimated an average groundwater flow velocity away from the river of 4.6 m/d. In contrast, Drury et al. (1984) used ^{14}C to estimate groundwater residence times of up to 15,000 y at distances of 50 km from the Murrumbidgee River, southeastern Australia, and calculated groundwater velocities of 10–50 m/y.

One of the best examples of the application of this technique to ephemeral rivers is the study in the Finke River, central Australia (Fulton, 2012). Carbon-14 ages in groundwater increase from less than 500 years within 3 km of the river to more than 12,000 y at distances beyond 40 km (Fig. 3). Based on the age gradient within 30 km of the river, the author calculated a groundwater velocity of 5 m/y. Assuming an aquifer thickness of 200 m and porosity of 0.22, he calculated a yearly mean recharge rate of $5–12 \times 10^6$ m^3 for a 35 km reach of the river.

One of the advantages of this method is that the technique does not require the researcher to be in the field at the time of the river flow event. It provides information on long-term fluxes, particularly when tracers such as ^{14}C are used. Of course, if the flow regime of the river has changed, then these fluxes might not be representative of current conditions. Groundwater velocities between about 5 and 500 m/y have been calculated using this technique, with higher velocities measured in perennial systems. Based on porosity of 0.1–0.4 and aquifer thickness of 10–100 m, this would correspond to infiltration rates of 100–10^5 m^2/y. Because ephemeral rivers only carry water for short periods of time, infiltration rates and groundwater recharge rates during flow events will be much greater than the long-term averages measured with many of these groundwater dating tools. Unlike many of the other techniques, this method provides information on actual groundwater recharge rates, rather than infiltration rates.

One of the difficulties of the technique is distinguishing between recharge from the ephemeral stream and recharge from other sources. This is often not of major concern in arid and semi-arid regions, where diffuse recharge is typically very low, and so river infiltration may provide the main source of groundwater recharge. However, it is still important to distinguish river recharge from regional groundwater flow, which originated upgradient of the river system. Uncertainties are also introduced in converting groundwater velocities to recharge rates. Although uncertainties in estimating groundwater age (and hence flow velocity) are often relatively small (approximately 10%), uncertainties in aquifer thickness and porosity can lead to uncertainties in fluxes on the order of ±50%.

3. Discussion

3.1. Choosing a method

One of the challenging and exciting aspects of ephemeral and intermittent stream research is that each stream presents unique and often unpredictable field conditions. Therefore, when choosing

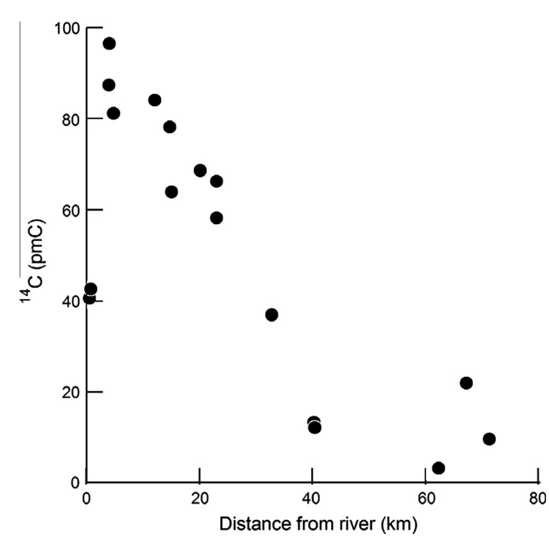

Fig. 3. Trend in carbon-14 activities in groundwater with distance from the Finke River, central Australia (Fulton, 2012). The decrease in carbon-14 activity reflects an increase in residence time, and can be used to determine the groundwater flow rate away from the river, and hence the recharge rate.

a method to estimate or measure groundwater recharge beneath ephemeral or intermittent streams, several factors must be taken into consideration:

(1) Spatial extent of the chosen study area. Methods that rely on measurements of unsaturated zone processes provide only point measurements of infiltration velocity. Significant work is required to extrapolate these measurements to the entire stream system. The hydraulic conductivity of streambed materials can range over more than eight orders of magnitude, and at individual field sites typically ranges over more than two orders of magnitude (Calver, 2001). Studies in gaining, perennial streams have found significant variations in rates of groundwater inflow over relatively small spatial scales, and similar results would be expected in losing, ephemeral streams (Cey et al., 1998; Kennedy et al., 2009). In comparison, groundwater based methods (groundwater mounding and chemical tracers) provide estimates of recharge with spatial scales of tens of meters to a few tens of kilometers, depending on the distance of the observation bores from the river (Fig. 4). Methods based on surface water flows (streamflow difference and floodwave front tracking) can provide estimates on spatial scales between several tens of meters and many tens of kilometers, depending on the distance between observation points.

(2) Time span of the study; whether the estimate is meant to capture the recharge resulting from individual events, the annual average, or a longer period. Methods that rely on measurement of unsaturated zone properties are frequently applied using sensors that can measure at short time intervals (i.e. seconds to minutes), and these methods also have a relatively fine temporal resolution. This can be important for understanding infiltration processes, and where information is required on the magnitude of infiltration from individual flow events. However, the length of record is limited by the duration of the study, and so would rarely extend to more than a few years (Fig. 4). The temporal resolution of streamflow-based methods can also be relatively short (i.e. minutes), but it is frequently possible to provide longer-term data if data from a gauging station with a long period of record (many years) is used. Temporal resolution of the groundwater mounding method depends upon the time for the groundwater mound to form and decay, which will be related to the hydraulic diffusivity of the aquifer and the distance between the streambed and water table. It will usually be on the order of a few hours to a few years. Chemical tracer methods provide an average recharge flux over periods from days to tens of thousands of years, depending upon the method used for determining the groundwater age.

(3) The range of recharge rates that can be reliably estimated by each method (Fig. 4). As discussed above, both unsaturated zone methods and streamflow based methods typically estimate infiltration rates associated with individual events. However, unsaturated zone methods typically estimate vertical velocities at a point. These values must be multiplied by the river width to obtain infiltration rates in terms of volume per length of stream per time (e.g., m^3/m/d or m^2/d). To convert infiltration rates measured during individual flow events into long-term averages, these values must be multiplied by the flow duration (e.g., hours per year or days per year). Thus, we might estimate an infiltration rate of 15 m^3/m for a flow event. Based on a flow duration of a few days, this would correspond to an instantaneous infiltration rate of 5 m^3/m/d. If flows of this magnitude occur twice per year, then this equates to an average infiltration rate of 30 m^3/m/y. For comparison between methods in Fig. 4, instantaneous flow velocities have been scaled assuming a stream width of 10 m, and these have been converted to long-term averages assuming a flow duration of three days per year. These values are considered typical for ephemeral streams. Since Fig. 4 only gives a broad indication of the likely range of fluxes that can be estimated using the different techniques in terms of order-of-magnitude, the exact

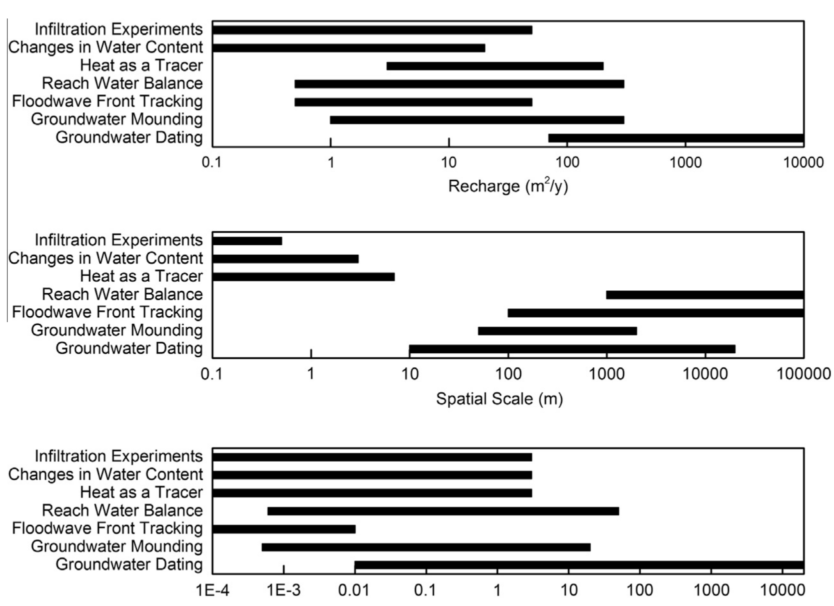

Fig. 4. Comparison of approximate spatial and temporal scales, and the range of recharge rates that can be reliably estimated with different methods (based on the studies cited in the text). Recharge rates given have been scaled to long-term average values, rather than instantaneous rates. Instantaneous rates have been converted to long-term average rates using a mean river width of 10 m, and flow duration of 3 days per year.

value used for this extrapolation is less important. We have attempted to give many quantitative examples of the ranges of recharge found in the literature in the previous sections.

(4) Equipment costs or availability and field time. Sometimes there is even a trade-off between the equipment and labor costs; for example, methods that use sensors with dataloggers may require more expensive equipment, but fewer visits to the (often remote) fieldsite. Methods that require groundwater levels or samples may be cost prohibitive where monitoring wells are not already existing in the necessary locations, especially where a large depth to groundwater requires deep wells.

(5) The presence of a clogging layer at the streambed surface. Because the spatial extent of this layer is commonly variable and transient, point estimates are not advised, and accurate determination of recharge requires either a time-integrated method such as groundwater mounding or groundwater dating, or continuous monitoring during a flow event or several events.

(6) Method uncertainty. Goodrich et al. (2004), provide a concise table of the trade-offs between data requirements, analyses, and uncertainties for several of the methods. Method uncertainty for an individual study may depend on several factors, including the limits of specific equipment, the distance and time between measurements, and what assumptions must be made about the system during data analysis.

3.2. Comparison of methods

Due to the different temporal and spatial scales of each measurement technique, and also because of their different strengths and weaknesses, it is always useful to estimate ephemeral river recharge rates using several methods. As described in Scanlon et al. (2002), this helps to constrain the errors and uncertainties associated with each individual method and to refine the conceptual model of recharge processes within the ephemeral system. It can be difficult to compare flood event estimates (i.e. from water content, heat tracer, water balance and floodwave front tracking) with temporally averaged methods (i.e. groundwater mounding or groundwater dating). Similarly, point estimates from tracer techniques are difficult to upscale for comparison with reach length values. However, these comparisons between different scales provide valuable information and overcome the disadvantages of each method.

For example, mountain front recharge can contribute a significant portion of basin aquifer replenishment in arid or semi-arid climates, and is typically estimated using chemical or isotopic tracers, rainfall-runoff models, streamflow gaging, or groundwater models (Wilson and Guan, 2004). However, spatial and temporal patterns in this recharge are typically poorly understood. By integrating floodwave tracking with heat tracer techniques, Niswonger et al. (2005) determined seepage losses along 15 km of intermittent, mountain front stream in Nevada, USA. Profiles of temperature measurements in three cross-sections of the streambed were used to estimate hydraulic conductivity and develop estimates of seepage losses based on stream depth. The authors found that, although the lower section of the stream only flowed during above average rainfall periods, the higher hydraulic conductivity in that section resulted in the contribution of a high percentage of total seepage when flow occurred.

As an example of comparison over different timescales, Drury et al. (1984) compared recharge rates through a perennial river in the Murray Basin in Australia using radiocarbon dating and hydraulic properties (Darcy's Law). The average recharge rate from ^{14}C ($0.016\ m\ d^{-1}$) was approximately half of that obtained from Darcy's Law. This agreement is probably reasonable considering the uncertainties of the two methods, particularly given the uncertainty in estimating hydraulic conductivity. Alternatively, the difference may indicate that recharge has decreased over time. Fulton (2012) extrapolated the recharge rate estimated from water table rise following a single flood to a mean annual recharge event by examining the frequency of flow events over the 24 year period for which river flow data was available. The calculated recharge rate of $5.7 \times 10^6\ m^3/y$ for a 35 km reach of the river was within the range of 5.1×10^6–$11.3 \times 10^6\ m^3/y$ calculated from environmental tracer (^{14}C) data. Agreement is good, particularly considering the different periods for which the estimates apply.

3.3. Infiltration versus recharge

Methods that rely on streambed properties or streamflow data usually provide information only on infiltration. Where the water table is deep, there can be significant delays between infiltration and groundwater recharge – sometimes up to tens of years (Jolly et al., 1989). Even when this time delay is accounted for, for a flow event of finite duration, recharge will always be less than infiltration. This is because after the stream dries up, some of the infiltrated water will evaporate. If the flow duration is very short (and the water table is deep), most of the infiltrated water may evaporate and so recharge from the flow event will be low or zero. The time delay between infiltration and recharge will depend upon the antecedent water content (the time delay will be less when the antecedent water content is greater), and hence so will the fraction of infiltration, which eventually becomes recharge. The proportion of infiltration which is lost to evapotranspiration will greatly increase if there are low permeability layers beneath the streambed, which reduce the infiltration rate.

As an illustration of this point, Shentsis et al. (1999) partitioned transmission losses into evaporation and recharge, by assuming the evaporative loss was initially equal to the potential evaporation rate and declined exponentially over time. The rate of decline was estimated based on field experiments, and allowed total evaporative loss to be estimated. They found that the evaporative loss was equal to the transmission loss for very small runoff events, but was insignificant for larger events. Overall, they found that evaporation was a small component of the transmission loss.

In contrast, a small fraction, if any, of the transmission losses recharge the regional groundwater in some systems. For example, evaporation or evapotranspiration is likely to be much greater than recharge along the Woodforde River in central Australia, even for high flow events. Although streambed infiltration during river flow events is very high, a low permeability layer occurs beneath the river sediments, which appears to reduce the infiltration rate and create a perched aquifer directly beneath the river. Recharge to the deeper regional aquifer is believed to be very low, with the regional groundwater not changing perceptibly in response to flood events (Fig. 5). The rate of water level decline in the perched aquifer is consistent with water use by large *Eucalyptus camaldulensis* trees (2.2–2.5 mm/d), which line the banks of the river (Cook et al., 2008). Similarly, a study on the distributaries of the Okavango Delta in Botswana concluded that infiltration from an intermittent stream generated a local freshwater lens in a generally saline environment where recharge was entirely consumed by trees and other vegetation and no connection to regional groundwater was established (Bauer et al., 2006). This situation can occur in parts of Australia and southern Africa, where old soils have developed impeding layers beneath channels, but may be less frequent in the arid basins of glaciated regions such as the western US or northern Africa, where soils are younger.

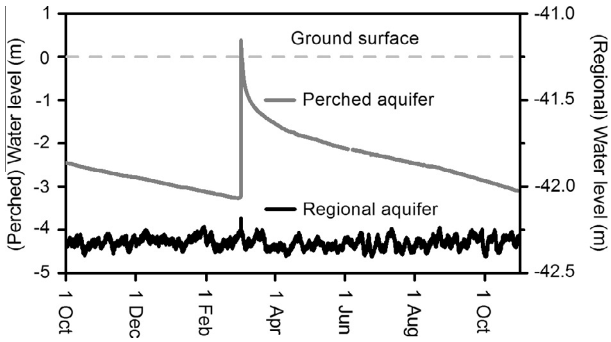

Fig. 5. Perched aquifer and regional aquifer response to a flood event in the Woodforde River, central Australia, during 2011–2012 (Villeneuve et al., in prep.). The rate of water loss from the perched aquifer is consistent with the rate of water use by river red gums (*Eucalyptus camaldulensis*), which line the banks of the river.

3.4. Change in infiltration over time

In most ephemeral stream systems, infiltration rates are not static over time, but can instead change between flood events or even within a single flood event, for example due to fine sediment deposition or changes in the water table. In a study in the Kuiseb Desert in Namibia, Dahan et al. (2008) identified three distinct ephemeral recharge processes: (1) the propagation of the wetting front into the soil, (2) steady infiltration to the water table while vadose zone water content stayed at field capacity, and (3) a significantly lower transmission loss once the water table rose to ground level. During the initial propagation of the wetting front into the soil, rates of infiltration in ephemeral streambeds can be extremely high, due to strong capillary pressure gradients (Blasch et al., 2006). These initially high, unsteady seepage rates can lead to challenges in estimating groundwater seepage using many methods, while providing an advantage for floodwave routing models (Niswonger et al., 2008). Several ponding experiments have clearly shown the transition between initial, capillary-driven seepage and later, steady-state infiltration. In a 7 m length of intermittent stream in southern Australia, Batlle-Aguilar and Cook (2012) observed a very significant transient infiltration effect, with high rates at the commencement of flow and again as the wetted perimeter of the stream increased for each increase in water depth. At a larger scale, Mihevc et al. (2001) flooded 60 m of an artificial channel in Nevada, USA for six days to understand longitudinal variability in seepage rates and subsequent areas of recharge to the local aquifer. As expected, the authors also found high seepage rates (>0.4 m/d) at the onset of the experiment due to the infiltration capacity of the initially dry soils. Once the experiment reached steady-state after a few days, volumetric seepage rates of 0.05–0.11 m/d were observed for a computed surface area of 477 m^2. The use of heat as a tracer during the experiment confirmed these seepage rates. In contrast, Freyberg (1983) showed that for very wide streams, the increase in flow width during rising hydrograph at start of flow can dominate transient infiltration effects, and mean that when averaged across the stream cross-section, infiltration rate increases over time. In any case, accurate characterization of streambed infiltration and aquifer recharge must account for both the highly variable infiltration at the onset of flooding, and the delay between initial infiltration and actual recharge due to changes in storage in the unsaturated zone.

Temporal variation in streambed temperatures can also cause transience in infiltration rates, because the density and dynamic viscosity of water increase as temperatures decrease. Using vertical measurements of temperature in the unsaturated zone beneath a pond, Jaynes (1990) observed reductions in infiltration as the surface water cooled diurnally. Similarly, Constantz et al. (1994) observed diurnal fluctuations in streamflow loss in streams in southwestern USA, and concluded that for diurnal changes in stream temperature greater than ~10–15 °C, transience in streamflow loss would be significant.

Finally, many authors describe the potential effects on infiltration due to lower permeability layers of fine sediment in the streambed. These "clogging" layers can develop during the flood, leading to lower infiltration rates over time, or be scoured from the streambed with the onset of flooding, resulting in higher transmission losses. Following a storm carrying very turbid, sediment-laden water into a channel in Colorado, USA, Susfalk et al. (2008) observed that seepage dropped below detection as the channel bed appeared to be effectively "sealed" by the deposited fines. During the following month changes in flow and relatively clear water resulted in the removal of this layer and the return of higher seepage rates (Fig. 6). Lange (2005) also observed fine layers of mud deposited at the streambed surface of the Kuiseb River during the recession of previous flood events, and supposed they may limit infiltration during small flood events that follow. Crerar et al. (1988) found that silt carried by floodwaters can effectively seal the surface of the streambed, greatly reducing the infiltration rate. This process occurred even at relatively high flow velocities, where sand grains were mobilised. In a study of the artificial recharge potential of wadis in Saudi Arabia, Missimer et al. (2012) observed that recharge from flood events is typically inefficient, partially due to the reduced rate of infiltration due to clogging from fine sediment loads. Some aquifer recharge enhancement schemes now consider the effect of fine sediment deposition on recharge rates. For example, the Omdel Dam in Namibia traps floodwaters and let the fines settle out before the resulting clear water proceeds through infiltration basins. This process has raised recharge after flood events from 20% to 50% of runoff (Seely et al., 2003).

4. Future directions

Overall, very few applied studies have been done to characterize recharge in these ephemeral and intermittent stream systems, leaving many research gaps. One of these gaps is actually the comparison of recharge estimates between methods. While some notable studies, as described in the discussion, have used multiple methods, there are relatively few such applied field studies. Even within these studies, the infiltration and recharge estimates are

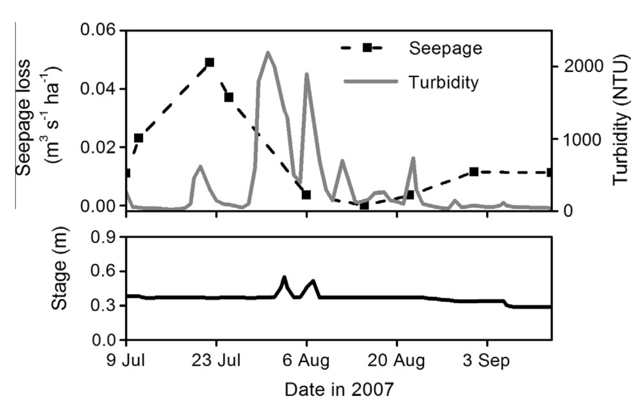

Fig. 6. Infiltration rates can change over time as variation in stage and turbidity deposits and scours fine material (i.e. "clogging layers") at the streambed surface. Transmission losses in the Rocky Ford Highline Canal in Colorado, USA, compared to changes in stage and turbidity. Following storms in July and August that carried turbid water with high loads of sediment from the Lower Arkansas River into the Rocky Ford Highline Canal in Colorado, USA, seepage losses dropped below detection. Subsequent scouring of the channel bed and relatively clear water slowly increased the infiltration rate in the months following the storm. Figure modified from Susfalk et al. (2008).

rarely reconciled, although the ultimate goal is generally to have a better understanding of aquifer recharge during flood events. In order to convert between infiltration or transmission loss and aquifer recharge, other factors such as the role of riparian vegetation must be known. The relative roles of bank and overbank infiltration during streamflow events would also need to be quantified, as transmission loss includes these processes, streambed infiltration does not, and aquifer recharge may result from these processes. Therefore, interdisciplinary studies linking the transmission losses (hydrology), infiltration (hydrology/soil physics), riparian response (ecology), and aquifer response (hydrogeology) would be extremely beneficial to the general, applied understanding of arid zone processes. Further, a comparative study to explore which methods likely underestimate or overestimate aquifer recharge, by how much, and due to which factors, would be highly valuable.

Another notable need is for more uncertainty analysis. To date, the published field studies usually apply a particular method (or less frequently, a combination of methods) and produce an estimate of transmission loss, infiltration, or recharge for a particular stream system. Rarely is the confidence in these estimates quantified. This is perhaps partly because many of the methods themselves are evolving, and many of the field studies provide a "proof of concept". Goodrich et al. (2004) took a notable first step towards this uncertainty analysis by providing a clear and useful table listing the most uncertain aspects of several methods used to determine recharge beneath a stream in Arizona, USA. Although the aspects listed were observations and therefore qualitative in nature, they could serve as a guide for future quantitative analysis.

Finally, like many other areas of science, arid zone aquifer recharge suffers from a general lack of long-term studies. Often, flood events occur between one and half a dozen times per year, leading to relatively small datasets temporally. The availability of multi-year studies would allow for an understanding of aquifer response to flood events over a wider range of both magnitude and duration. This increased knowledge would also help in the prediction of the possible effects of climate change on arid zone aquifer levels.

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