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Understanding and quantifying focused, indirect groundwater recharge from ephemeral streams using water table fluctuations

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1	Understanding	and	quantifying	focused,	indirect	groundwater	recharge	from
2	ephemeral streams using water table fluctuations							

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14

15 Key Points

- A water table fluctuation method to quantify indirect recharge is presented
- 17 Indirect recharge decreases almost linearly away from a semi-arid mountain front
- This spatial pattern is persistent both in the long term and on an event basis

19

20 Keywords

- 21 Groundwater recharge, ephemeral stream, water table fluctuation, semi-arid hydrology,
- 22 indirect recharge, focused recharge, dryland hydrology, mountain front

23 Abstract

Understanding and managing groundwater resources in drylands is a challenging task, but 24 one that is globally important. The dominant process for dryland groundwater recharge is 25 26 thought to be as focused, indirect recharge from ephemeral stream losses. However, there is a global paucity of data for understanding and quantifying this process and transferable 27 techniques for quantifying groundwater recharge in such contexts are lacking. Here we 28 develop a generalised conceptual model for understanding water table and groundwater head 29 fluctuations due to recharge from episodic events within ephemeral streams. By accounting 30 for the recession characteristics of a groundwater hydrograph, we present a simple but 31 powerful new water table fluctuation approach to quantifying focused, indirect recharge over 32 both long term and event timescales. The technique is demonstrated using a new, and 33 globally unparalleled, set of groundwater observations from an ephemeral stream catchment 34 35 located in NSW, Australia. We find that, following episodic streamflow events down a predominantly dry channel system, groundwater head fluctuations are controlled by pressure 36 37 redistribution operating at three timescales from vertical flow (days to weeks), transverse 38 flow perpendicular to the stream (weeks to months) and longitudinal flow parallel to the stream (years to decades). In relative terms, indirect recharge decreases almost linearly away 39 from the mountain front, both in discrete monitored events as well as in the long term 40 average. In absolute terms, the estimated indirect recharge varies from 80 to 30 mm/a with 41 the main uncertainty in these values stemming from uncertainty in the catchment scale 42 hydraulic properties. 43

45 **1. Introduction**

46 Dryland regions (semi-arid and arid regions but excluding hyper-arid deserts) are expanding and now represent ~35% of the global landmass, support a population of around 2 billion 47 48 people (90% of which live in developing countries), 50% of the world's livestock, 44% of all cultivated land and contain some of the most important wetlands in the world [Hassan et al., 49 2005]. Water scarcity is becoming more critical in dryland areas due to population growth 50 51 and urbanisation, increasing irrigation demands and climate change [Scanlon et al., 2006; Taylor et al., 2013]. In the wider Earth Science context, understanding groundwater recharge 52 processes in drylands is also important for the interpretation of paleoclimatic proxy archives 53 54 [Cuthbert et al., 2014], and their longer term sensitivity to change. Furthermore, understanding the relationships between climate and groundwater availability in drylands 55 may enable us to understand better our own origins as human beings [*Cuthbert and Ashley*, 56 57 2014]. However, the understanding and quantification of groundwater recharge processes in dryland areas remains a major challenge worldwide [Wheater et al., 2010]. 58

59 In drylands the climate has large atmospheric water demands and temperature contrasts, surface water flows are infrequent but potentially damaging and populations are sparse and 60 often have limited economic resources [Wheater et al., 2010]. Groundwater recharge in 61 drylands predominantly occurs via leakage from ephemeral streams [Simmers, 1997; 2003]. 62 Recharge can also occur more diffusely under the right conditions. For example where 63 sufficient preferential flow pathways exist to enable flow to by-pass otherwise dry soil 64 profiles, or where soil moisture deficits are limited due to thin soils or lack of vegetation 65 [Cuthbert and Tindimugaya, 2010; Cuthbert et al., 2013], or in Mediterranean climates with a 66 winter rainy season when evapotranspirative losses are lower [van Loon and van Lanen, 67 2013]. However, these diffuse processes are, arguably, more widely understood and already 68 69 successfully included in large scale hydrological models, while the major areas of uncertainty exist in areas where recharge from surface-water bodies such as ephemeral streams
dominates [*Döll and Fiedler*, 2007; *Epstein et al.*, 2010; *Wheater et al.*, 2010]. Following
Healy [2010] here we use the term 'focused recharge' to refer to any recharge from a surface
water body, and 'indirect recharge' as a sub-type of focused recharge whereby recharge
occurs due to infiltration from streambeds such as the ephemeral streams that drain semi-arid
mountain front systems.

Systematic, multi-year observations of groundwater dynamics in ephemeral stream 76 catchments are very rare and only reported for a few sites worldwide [Besbes et al., 1978; 77 Carling et al., 2012; Goodrich et al., 2004; Pool, 2005; Shentsis and Rosenthal, 2003]. Most 78 79 dryland hydrological studies have been 'top down', attempting to characterise groundwater recharge using a water balance approach based on surface measurements. Such methods are 80 complicated by the inherent non-linearities in predicting rainfall-runoff relationships, the 81 difficulties of measuring flows and therefore transmission losses accurately in such 82 environments, and transience in the nature of streambed losses [Shanafield and Cook, 2014]. 83 84 Where transmission losses can be measured well or predicted, estimations of recharge are then hampered by the difficulty of estimating transpiration losses and/or lateral subsurface 85 86 flow behaviour due to alluvial structures [*Telvari et al.*, 1998]. Furthermore, upscaling from 87 point scale measurements to larger scales can be highly problematic [McCallum et al., 2014].

In contrast, observations of the water table fluctuations of a catchment can provide the most direct measure possible of the recharge behaviour, as they integrate the recharge response over a spatial footprint much larger than that of the measurement (borehole) scale. Estimating indirect recharge from time series of groundwater level measurements has been the subject of much research, but almost exclusively focused on inverse solutions of the transient mounding equations in various forms [*Abdulrazzak and Morel-Seytoux*, 1983; *Dillon and Liggett*, 1983; *Hantush*, 1967; *Moench and Kisiel*, 1970] However, this previous work has not generally 95 accounted for the background groundwater recession behaviour or lateral boundary 96 conditions. Furthermore, published studies are mostly based on data from a single piezometer 97 or single event, therefore restricting its applicability. Finally, the available analytical 98 approaches struggle with the complexity of the form of the input function for time varying 99 recharge.

In this paper we first develop a generalised conceptual model for understanding water table fluctuations in ephemeral stream catchments using insights gained from analytical and numerical models of idealised aquifers. By accounting for the recession characteristics of a groundwater hydrograph we then present a simple but powerful new approach to quantifying indirect recharge separately over both the long term and on an event basis. This model is then tested using a unique monitoring database of groundwater dynamics from an ephemeral stream catchment in NSW, Australia.

107 2. Theoretical Background

108 **2.1** The water table fluctuation method for quantifying recharge

109 The basis of the water table fluctuation (WTF) technique for quantifying recharge is the110 following equation:

111
$$R = S_y \frac{\partial h}{\partial t} + D \tag{1}$$

where *R* is the rate of recharge $[LT^{-1}]$, S_y is specific yield [-], *t* is time [T], *h* is hydraulic head [L] and *D* is the rate of net groundwater drainage (or 'rate of groundwater flux recession') [LT^{-1}] [*Cuthbert*, 2010]. This assumes that changes in groundwater level in an aquifer are controlled solely by the balance of recharge and net groundwater drainage away from a given observation point and ignores other factors such as entrapped air, barometric fluctuations or local groundwater abstraction. The main limitations of the WTF method stem from difficulties of defining and estimating specific yield, and accounting for the drainage term (*D*)
robustly [*Healy and Cook*, 2002].

2.2 The general form of water table fluctuations in catchments dominated by indirect episodic recharge

An improved understanding and estimation of D has recently been proposed for the 1-D 122 groundwater flow equations under uniform recharge [Cuthbert, 2010; 2014]. However, an 123 adequate method for dealing with the recessional characteristics of a catchment in which 124 recharge is dominated by losses from an ephemeral stream has not so far been proposed. It is 125 therefore addressed here with regard to the idealised 2-dimensional aquifer shown in 126 Figure 1. It is bounded at one end (at x = L) by a no flow boundary – this may represent the 127 edge of an alluvial aquifer abutting a mountain front for example, typical in headwater 128 129 ephemeral stream settings [Pool, 2005; Simmers, 1997]. The aquifer episodically receives surface runoff via a stream channel flowing in the x-direction from higher elevations across 130 131 this boundary which is then received by the aquifer beneath via streambed infiltration during episodic flow events. The downstream boundary condition at (x = 0) is a constant head 132 boundary representing a typical discharge zone such as the transition to a perennial stream, 133 wetland or terminal lake. The lateral boundaries are no flow, thus the system is representative 134 of a series of parallel ephemeral streams, again a reasonable simplification in a dryland 135 setting. The linearised groundwater flow equation in 2-dimensions for such an aquifer, here 136 assumed to be homogeneous and isotropic, may be written as follows: 137

138
$$R = S_y \frac{\partial h}{\partial t} - T\left(\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2}\right)$$
(2)

where *T* is transmissivity $[L^2T^{-1}]$, and *x* and *y* are orthogonal length variables [L] as shown in Figure 1. This linearisation assumes that the fluctuations in water table elevations are small compared with the saturated thickness of the aquifer. 142 For some time during and after an episodic streamflow event we would expect a groundwater mound to rise and decay in the vicinity of the stream. Assuming that the recharge occurs 143 along the length of the stream, it is effectively acting as a line source during the recharge 144 period. We would thus expect the pressure wave generated to propagate transversely towards 145 the lateral boundaries, at a distance W in the direction perpendicular to the stream, with an 146 aquifer response time (ART), or time constant, of $t_{lat} = W^2 S_y / T$ [Currell et al., 2014; 147 Domenico and Schwartz, 1998; Rousseau-Gueutin et al., 2013]. This aquifer event response 148 will be superimposed on a longer term background recession acting longitudinally in the 149 150 direction parallel to the stream due to drainage to the perennial stream reach downstream, with a characteristic ART of $t_{long} = L^2 S_y / T$. 151

152 It is clear from a comparison of Equations 1 and 2 that the groundwater flux recession rate,153 *D*, is given by:

154
$$D = -T\left(\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2}\right)$$
(3)

155 The first and second terms on the RHS of Equation 3 express the superposition of the 156 longitudinal recession and the transverse recession respectively.

To illustrate these concepts the scenario described above and illustrated in Figure 1 has been 157 modelled numerically using COMSOL Multiphysics (v5.1). The indirect recharge was 158 simulated as an imposed flux boundary condition across a constant width of 20 m. This 159 implicitly assumes that there is insignificant lateral spreading of the wetting front beneath the 160 stream which is reasonable for cases where the depth to water table is less than the width of 161 the channel [Nimmo et al., 2002]. However the applied recharge from the channel varied in 162 163 space along the reach, with recharge decreasing linearly to zero between the upstream and downstream boundaries - an arbitrary distribution but one which mirrors the finding of 164 previous research, that indirect recharge decreases away from runoff source areas such as 165

mountain blocks [*Simmers*, 1997]. A long time series of identical episodic recharge events,
each with a constant flux and duration, was modelled to bring the system to a quasi-steady
state. The heads at points 1-4 were then output from the model for the last event and are
shown as hydrographs in Figure 2. The parameters used are given in the legend for Figure 1.

170 Figure 2 shows how the background (longitudinal) recession is expressed as a straight line with a transverse mounding event superimposed upon it. The timescale for the decay of the 171 mound can be estimated using an analytical solution. The analogous idealised problem of the 172 173 1-D redistribution of heads following a change in flux at one boundary (i.e. y = 0 at the stream), and a no flow boundary at y = W (i.e. an aquifer half space assuming parallel 174 streams) is given by Bruggeman's Equation 135.02 [Bruggeman, 1999]. Using this solution, it 175 is possible to show that 99% of the transience created by a change in flux at the stream 176 boundary will have decayed away within $t = t_{mound} \sim W^2 S_y/(2T)$ (i.e. half of $t_{lat.}$) since the 177 178 change in flux. For the present case of the ideal aquifer example plotted in Figure 2, $t_{mound} \sim 100$ days. 179

Furthermore, where recharge is distributed evenly across a catchment, recent theoretical work 180 [Cuthbert, 2014] shows that straight line recession behaviour is expected prior to 181 $t_{lin} = x^2 S_{v}/(16T)$ since a recharge event occurred, where x is the distance from the monitoring 182 point to the downstream fixed head boundary. In our modelled example, L is significantly 183 greater than W, as you would expect in most natural settings, and thus t_{mound} is smaller than t_{lin} 184 over much of the catchment. Hence, the straight line recession is observable under such 185 conditions, as long as the time between recharge events is greater than t_{mound} . A further point 186 187 worth noting here is that straight line recessions are also expected in contexts where flow lines are divergent [Cuthbert, 2014]. Thus where an aquifer is bounded by streams that are 188 189 not parallel, the mounding timescales may vary along the length of the streams, but the long

term recession would still be expected to be linear at early times following the cessation ofrecharge.

Straight line background recessions are observed in our synthetic example in line with the theory developed for evenly distributed recharge, despite the modelled recharge actually varying spatially. It is important to demonstrate that this feature of longitudinal recessions is a generally applicable one for the case of spatially variable recharge. Thus, additional analysis is needed as outlined in the next section.

197 2.3 Groundwater flux recession in catchments with spatially variable recharge

An expression for the recession of an ideal 1-D aquifer from an arbitrary initial condition is given by equation (10) of Venetis [*Venetis*, 1971]. In order to test the possible form of the longitudinal recession for the case considered above (i.e. recharge increasing linearly from zero at a downstream constant head boundary condition (h = 0) at x = 0 to R_{max} at x = L) it is useful to set the initial condition ($h_0(x)$) to the head distribution under steady state conditions. For $R = R_{max}x/L$, then it is straightforward to show that:

204
$$h_0(x) = -\frac{R_{max}x^3}{6LT} + \frac{R_{max}Lx}{2T}$$
 (4)

205 Venetis (1971, equation 10) gives the following expression for the variation in head as:

206
$$h(x,t) = \frac{1}{L} \sum_{n=1,3,5...} e^{\frac{-n^2 \pi^2 T t}{4SL^2}} \sin\left(\frac{n\pi x}{2L}\right) \int_0^{2L} h_0 \sin\left(\frac{n\pi x}{2L}\right) dx$$
(5)

From this equation, the conditions under which spatially variable recharge should produce straight line recessions can be analysed. Since we are only considering 1-D (longitudinal) flow in this case, (i.e. just considering the recession which occurs after any mounding due to indirect recharge, and variation in head in the y-direction, has dissipated) the net groundwater drainage can be simplified to:

$$D = -T\frac{d^2h}{dx^2} \tag{6}$$

Equations 4 to 6 have been used to plot Figure 3 with D normalised to the recharge value at the mid-point of the model domain (x/L=0.5). This shows how the modelled groundwater flux recession rate varies following a recharge event relative to the initial recharge rate across a range of ARTs. Values close to 1 on the vertical axis thus indicate that the recession is a straight line and accurately predicts the spatially varying recharge rate.

This shows that for some time following cessation of recharge, the straight line recessions are a direct indicator of the variation of the spatial variability in long term recharge. Furthermore, this analysis indicates that even for this case of spatially varying recharge, $t < t_{lin}$ [*Cuthbert*, 2014] can provide a reasonable (and conservative) measure of the length of time straight line recessions can be expected to last.

223 2.4 A water table fluctuation method for quantifying indirect recharge

Based on the preceding theory, we can now propose a new WTF approach to estimating 224 episodic indirect recharge. As with any WTF recharge estimation methodology, it should 225 226 only be used if a robust conceptual model warrants it. Thus, as per the methodology outlined by [Cuthbert, 2010] for estimating diffuse recharge using WTFs, the first steps to be taken 227 should be delineating the main hydrogeological boundaries, considering the likely controls on 228 recharge due to the presence of superficial deposits and the climatic context, and utilising 229 estimations of aquifer properties where possible. Furthermore, the time series of groundwater 230 level data to be used must be of sufficient temporal resolution, representative of the local 231 water table position, and sufficiently distant from the influence of pumping wells. 232

The analytical and numerical models of an idealised catchment described above, suggest that where a straight line groundwater level recession is observable, it can be used in two ways to estimate indirect groundwater recharge:

1. The slope of the straight line recession can be used to estimate the 'long term' ratio of R/S_y (or the actual recharge if S_y is known) by the following equation:

238
$$R_{av} = S_y \frac{\partial h}{\partial t}$$
(7)

Since the antecedent history of the system is not necessarily known, the meaning of 'long term' cannot always be precisely determined. However, as Figure 4 indicates, away from the fixed head boundary, the aquifer damps out variations in recharge so that significant variations in flux recession rate only occur due to recharge variations with periods less than the ART. Hence, away from a fixed head boundary, observation of a straight line recession and use of Equation 7 will provide an estimate of the recharge occurring over the previous time period defined by the ART (Figure 4).

246 2. On an event basis, the background recession can be added to a groundwater hydrograph 247 time series to reveal the change in head due exclusively to event recharge from the stream. 248 This is illustrated in Figure 5, where the effect of the long term recession rates have been 249 removed in this way from the groundwater hydrographs already shown in Figure 2. If the 250 system is behaving in the manner expected by the conceptual model outlined, for $t > t_{mound}$, 251 the result should be a step change in head (Δh) where:

Figure 5 indicates that, with the longitudinal recession removed, significant head increases still occur nearer to the stream due to the transverse spreading of the pressure wave generated by the flow event (hydrographs 1 and 3). However, further away from the stream (hydrographs 2 and 4) this effect becomes almost unnoticeable, with the response nowresembling a gradual step change in head.

Both techniques ultimately rely on knowing the value of S_y for estimating actual recharge and this can be challenging to obtain at the right spatial scale. However, S_y can be estimated from the definition of t_{mound} ($W^2S_y/(2T)$) if t_{mound} is determined by observation, *T* is estimated for example from a pumping test, and *W* from the geometry of the system.

262 3. A case study from Middle Creek, NSW, Australia

263 **3.1 Catchment context**

The catchment has been described in detail previously [Andersen and Acworth, 2009; Rau et 264 al., 2010] and is only briefly summarised here. Middle Creek (via Horsearm Creek) is an 265 ephemeral tributary to Maules Creek, itself a tributary to the Namoi river in the headwaters of 266 the Murray Darling Basin, NSW, Australia (Fig. 6). The Nandewar Range (part of the Great 267 268 Dividing Range) to the north-east receives approximately 1100 mm per year of precipitation in the long term. Rainfall is generally well distributed throughout the year, however the 269 rainfall intensity varies substantially with heavy rains generally occurring in the summer 270 months (Dec – Feb). The rainfall is also influenced by longer term fluctuations in the El Nino 271 Southern Oscillation Index (ENSO), with higher than average rainfalls in the positive phase 272 (La Nina) and lower than average rainfalls in the negative phase (El Nino). 273

Large storm events generate runoff from the steep headwaters of Middle Creek catchment which is comprised of Miocene volcanic rocks overlain by thin soils with forested land use. Flow is delivered across the mountain front (defined by a thrust fault) and onto a moderate gradient (1 to 2%), Quaternary age, alluvial fan up to 40 m thick. This overlies Permian sedimentary deposits (claystones, siltstones, sandstones, conglomerates and coal measures) 279 and Carboniferous crystalline rocks, meta-sediments and volcanic deposits. The degree of hydraulic connectivity between the Quaternary alluvium and these underlying formations is 280 presently unknown. As can be seen in Figure 6, the land downstream of the mountain front is 281 282 largely cleared for grazing, except for a narrow vegetated zone adjacent to the creek. A well delineated ephemeral channel has cut through clay rich soils which otherwise blanket the 283 alluvium. The main channel is typical of an episodic high energy stream comprising sand and 284 285 gravel deposits often forming pool-riffle sequences and cobble to boulder size lag. Ephemeral flows have been observed to extend all the way to the confluence with Horsearm Creek and 286 287 Maules Creek. Rainfall on the alluvial fan itself decreases to the southwest away from the Nandewar range. At Middle Creek Farm, the recent record indicated 522 mm/a for 2014, 288 Bellevue farm situated further downstream averages 534 mm/a, and both are in contrast to the 289 290 912 mm/a for Mt Kaputar in the catchment headwaters (see Figure 6 for locations).

For the time series available, Middle Creek flows when the cumulative rainfall in the month prior exceeds around 140 mm and the majority of runoff is assumed to be generated in the steep and low permeability mountain headwaters. The regional hydraulic gradient indicated by available groundwater level data is approximately northeast to southwest. There is little groundwater pumping in the Middle Creek area itself. However Middle Creek is just one of a series of ephemeral streams draining into Maules Creek and providing recharge to aquifers which are extensively pumped for cotton irrigation in the Namoi valley downstream.

298 **3.2** Monitoring installations & testing methods

Six 0.168 m diameter boreholes (Fig. 6) were drilled in 2012 using an air flush rotary/hammer method with advancing steel casing and installed with either two or four multilevel piezometers, each screen being hydraulically isolated using bentonite seals. After completion, the piezometers were developed by air-lifting using a compressor. Care was taken not to blow air into the screened section of the piezometers. Air-lifting was continued until the discharged water was clear. Details of the resulting 20 piezometers are given in the Supporting Information. The drill cuttings revealed that the alluvium comprises a highly heterogeneous layered system of mixed gravel, sand, clay and silt. The large variation in grainsize is as expected given the alluvial fan depositional setting.

Every piezometer was monitored at 15 minute frequency using Solinst Leveloggers, 308 compensated using a barometric logger situated in a borehole at East Lynne (BH20) which 309 recorded air pressure at exactly the same times. This was hung in the piezometer at 310 approximately 2 mbgl to avoid large temperature variations and thereby minimise any diel 311 artefacts in the pressure data [Acworth et al., 2014], whilst remaining above the water 312 column. Manual dip-tape measurements were made each time the data were downloaded and 313 used to check no significant drift in the loggers was occurring. These measurements were 314 315 also used, in combination with elevation data of each borehole datum measured using a Differential GPS, to convert the data to hydraulic head with respect to Australian Height 316 317 Datum (AHD).

A constant rate pumping test was carried out at Elfin Crossing (BH14, Fig. 6) and analysed using a transient model. For this analysis the Theis [*Theis*, 1935] equation was used incorporating the superposition of an injection image well to implement a recharge boundary due to the close proximity of the perennial section of Maules Creek (~35 m). The drawdown data were fitted to the model by varying the hydraulic parameters (*T*, *S*) in order to minimise the RMSE. Details of the pumping test and analysis can be found in the Supporting Information.

325 Stream stage was measured adjacent to Boreholes BH20-BH21 at East Lynne using a 326 Campbell CS450 pressure transducer logged by a Campbell CR1000 since June 2013. Since

327 June 2012, a digital camera placed at East Lynne has been capturing a record of flows in the creek which can also be used to determine the timing and approximate magnitude of the flow 328 events. It is noted that some small but greater than zero stage measurements are apparent 329 330 between Sep 2013 and March 2014 in the East Lynne stage hydrograph (Fig 7) caused by temperature driven air pressure differences between the transducer in the creek and the hut on 331 the creek-bank in which the data logger was housed. However, based on site visits and 332 333 photographic evidence from the automated on-site camera, there was no flow in the creek during this period. 334

A full Campbell weather station was installed next to BH19 at Middle Creek Farm and hasbeen recording since August 2013.

337 3.3 Groundwater hydrograph dynamics

338 Time series of heads recorded in every piezometer is shown in Figure 7 alongside the stream hydrograph at BH20 (East Lynne) and the cumulative rainfall record from Mt Kaputar. We 339 consider this to be a globally unparalleled dataset with respect to the intensity of groundwater 340 level data being collected in an ephemeral stream catchment, allowing an unprecedented 341 insight into its hydrodynamics. More detailed additional plots for nearby groups of 342 piezometers are given in the Supporting Information Figure S1. Heads varied between 343 3 and 8 m below ground level with greatest unsaturated zone thickness occurring beneath the 344 345 streambed at the most upstream location (BH20 and BH21), and the greatest total unsaturated 346 zone thickness occurring at the top end of the reach, furthest away from ephemeral streams 347 (BH22). There is evidence of barometric fluctuations seen in piezometers from BH18-20, but not in BH17. This suggests the presence of materials with low permeability above the 348 349 screened depth in the BH18-20 piezometers [Acworth et al., 2014]. Loading responses also occur at times of episodic surface flows as indicated by sudden increases in head seen in the 350

groundwater hydrographs, corresponding with sudden stream stage increases in Middle 351 Creek. This is consistent with the variable lithology encountered during drilling, and the 352 variability in formation hydraulic conductivity implied by drawdowns observed during 353 354 hydrochemical sampling. However, in general the groundwater fluctuations are dominated by increases coincident with stream flow events in Middle Creek followed by recessions. The 355 exceptions are BH20_1 and BH22_4 which are clearly screened within low permeability 356 357 units and therefore show very slow responses in comparison to the other piezometers. Following an ephemeral flow event, groundwater head changes are characterised by a rapid 358 359 increase in gradients between piezometers followed by a more gradual re-equilibration occurring on three distinct length and time-scales of hydraulic head redistribution. These can 360 be interpreted as being due to vertical, transverse and longitudinal propagation of the pressure 361 362 increase induced by stream flow losses to the underlying alluvium. Vertical downward hydraulic gradients are initially induced near the creek which then dissipate on the timescale 363 of days to weeks (for example compare BH17_1 and BH17_4 in Figure 7). Transverse 364 gradients away from the creek dissipate on the timescale of weeks to months, and 365 longitudinal, down-catchment, gradients are apparent throughout the whole monitoring 366 period suggesting they persist over longer timescales of years. 367

Consistent with the idealised groundwater hydrograph responses to episodic indirect recharge 368 described in Section 2 (Figure 1 & 2) there is an observed time lag and amplitude attenuation 369 with distance away from Middle Creek which is particularly pronounced at the most distant 370 location from Middle Creek, BH22. Also akin to the idealised hydrographs is the mounding 371 which occurs after a stream flow event, followed by a gradual transition to a straight line 372 recession during extended periods of no stream flow. In this case t_{mound} , estimated as the time 373 between the cessation of surface flow in the creek and the return to conditions of straight line 374 groundwater recession, is approximately 135 days. Thus the straight line recession is only 375

seen once in the time series (Feb-Mar 2014) when the time between stream flow events exceeds this timescale. As shown by the dashed black lines in Figure 7, the steepness of these long term recessions decreases with distance downstream. For BH20, BH21 & BH22, located a similar distance from the mountain front, but different distances from Middle Creek (2 m, 37 m and 1111 m, respectively) the mounding behaviour is initially different for each borehole. However, at later times the recession converges to a remarkably consistent straight line gradient, as expected from the theoretical considerations discussed above.

It is noted that during streamflow events, the head in BH21 rises to a slightly lower absolute 383 level than the head at BH20, and there is flow away from the creek for the duration of the 384 385 flow event, consistent with our conceptual model. However, the recession at BH21 is larger than that of BH20 which we believe is due to pumping from a nearby stock watering well 386 located around 60 m west of BH21. It is a very minor abstraction (intermittent, wind-mill 387 388 driven pump and the numbers of livestock observed in the vicinity are low) although it nevertheless appears to produce a local cone of depression which influences the variation in 389 390 water level at BH21. This is discussed further in section 3.4 in terms of its implications for 391 the estimation of recharge. We also note that groundwater use by riparian vegetation remains 392 unknown at this site. However such water use is certain to contain a significant soil moisture 393 component, especially during and following recharge events. Any impacts on the water table by direct groundwater use are therefore likely to be very small relative to the broader trends 394 observed. 395

Small vertical head gradients, developed in response to streamflow events, are observed in the logger data at some locations and these differences are consistent with manual dip-tape measurements, thus not being artefacts of the logged time series. For BH17 and BH20 immediately adjacent to the stream, downward gradients occur after surface water flow events, consistent with the indirect recharge mechanism proposed, which then dissipate

401 quickly over time since the event. In contrast at BH19, situated 50 m away from the stream, the gradient is upwards following streamflow events. This is suggestive of water propagating 402 through a more permeable layer at depth whilst equilibrating vertically as the recharge pulse 403 404 dissipates transversely. Since the small vertical gradients equilibrate on a timescale much shorter than the transverse head gradients, it is a reasonable assumption that the groundwater 405 level observations are mostly representative of the water table dynamics during the transverse 406 407 and longitudinal recession periods. Thus it is reasonable to apply the methodology proposed in Section 2 which was based on a 2-D representation of an idealised aquifer which assumes 408 409 no vertical flow is occurring.

410 **3.4** Quantifying recharge using the new methodology

411 The long term straight line recessions were calculated using the data shown in Figure 7. A complication in this task was that each borehole hydrograph has a varying degree of 412 barometric 'noise' in the water level signal. Thus, a purely statistical approach, for example 413 414 using a cut-off for a particular coefficient of determination on a linear regression, was not 415 deemed appropriate. Our approach was, therefore, to identify the time period from which the hydrographs at different distances from the creek converged onto a consistent linear recession 416 417 after a stream flow event and until the groundwater levels began to respond to the next stream flow event. Since this is somewhat subjective, two reasonable end member times were 418 selected for start of the linear recession period, in this instance 2 weeks apart from each other, 419 in order to account for the subjective uncertainty (further apart than this and the mismatch 420 becomes obvious). The recession rates were then calculated by averaging the incremental 421 422 changes in head during the assigned periods. These two recession rates were then 'removed' from the head time series by adding the calculated values, and the average residual heads 423 have been plotted in Figure 8. The event based estimates of R/S_v summed over 2013 were 424 425 then calculated from Figure 8 and plotted against the long term estimates in Figure 9, with error bars added to indicate the uncertainty in the analysis due to the variation in the chosenrecession rate. A summary of these values and their uncertainties is given in Table 1.

It is noted that the residual head increase at BH21 is larger than that at nearby BH20, probably due to the minor nearby abstraction as discussed in section 3.3. However, the long term recession from BH21 eventually begins to converge with those of BH20 and BH22 suggesting that the effect is a transient one which diminishes during dry periods. Hence, although the derived R/S_y values for BH21 are likely to be overestimates they are still within the error bounds for BH20 and BH22.

In order to convert the estimates of R/S_v presented above into actual recharge, S_v must first be 434 estimated. A best fit ($R^2=0.99$) value for T from the pumping test on BH14 was 115 m²/d (see 435 436 Supporting Information). From Figure 7, the mounding timescale at East Lynne can be estimated as the time from the cessation of flow in the stream until the convergence of the 437 recessions onto a straight line, which is 135 d with an uncertainty of +/- 7 days as previously 438 439 assigned to account for the uncertainty in the choice of the start of the straight line recession 440 period. The half-space, W, can be estimated by halving the average distance from Middle Creek to the adjacent ephemeral creeks. Since there is some convergence of the adjacent 441 creeks, and therefore variation in W, with longitudinal distance downstream from the 442 mountain front (Figure 6), this calculation was only done for the East Lynne location where 443 the adjacent streams are close to being parallel. Allowing for some uncertainty due to the 444 slight convergence in the streams we estimate W to be 1.6 km + - 0.1 km. Using the 445 expression for t_{mound} given in Figure 2 this implies that the diffusivity (T/S_y) is ~9500 +/- 700 446 m²/d. Taking the pumping test value for T (115 +/-12 m²/d), yields a value of S_v of 0.012 +/-447 0.003. This is reasonable given the prevalence of interbedded clay layers and also a 448 significant proportion of fines within many layers of the alluvial material encountered in the 449 450 catchment. Since a stage hydrograph was only available at East Lynne, t_{mound} could only be

451 estimated for this location, but the resulting S_y value was applied to all piezometers. A 452 summary of the recharge estimates and their uncertainties is given in Table 1.

What is immediately apparent is that, assuming S_{ν} is not varying significantly within the 453 catchment, the amount of groundwater recharge is generally decreasing with increasing 454 distance from the mountain front. Furthermore, this trend is consistent between the long term 455 and event based estimates suggesting that this is a persistent feature of the recharge behaviour 456 in the catchment. Groundwater recharge for 2013 was lower than the estimated long term 457 average by around 23% (Figure 9). The long term average value is representative of recharge 458 occurring over the preceding period given by the ART which, using the above values for the 459 460 catchment hydraulic diffusivity and a length of 10 km, is approximately 30 years. Using the estimate for S_y of 1.2% enables us to estimate the long term (30 year) recharge in the 461 catchment using this technique as over 70 mm/a close to the mountain front (BH20-22) and 462 463 around 30 mm/a by 6 km further downstream. Similarly, indirect recharge for 2013 has been calculated and plotted against distance from the perennial downstream boundary indicating 464 an almost linear relationship (Figure 10). The zero recharge point is defined as the most 465 upstream perennial section of Horsearm Creek, 2 km upstream of Elfin Crossing. As a reality 466 check for this system, since the change of recharge with longitudinal distance along the 467 stream appears to be approximately linear, we have applied equation 4 with $R_{max} = 68$ mm/a, 468 the maximum estimated for 2013, $T = 115 \text{ m}^2/\text{d}$ and L = 10000 m. The computed and 469 observed heads during a recession period in 2013 are plotted in Figure 10. The comparison is 470 good given the simplicity of the model, and demonstrates further consistency between the 471 derived recharge values, estimated aquifer parameters and the groundwater observations. 472 While we acknowledge the uncertainty in the absolute magnitude in the recharge estimations, 473 474 this highly heterogeneous alluvial system is a very challenging one in which to estimate

475 hydraulic properties at the appropriate scale and in more homogeneous aquifers, the 476 estimation of *T* or S_y should be even more straightforward.

477 **3.5.** Deviations between real and ideal catchment behaviour

Although the methodology we have presented is potentially very powerful, as with most 478 analytical methods, several issues arise when applying them to field conditions. For instance, 479 the model assumes parallel adjacent ephemeral channels but the field example includes 480 adjacent channels that converge within the study reach. As noted, such deviation in the 481 geometry will affect the accuracy of the t_{mound} estimations for calculating hydraulic 482 parameters. However, straight line recessions are theoretically predicted [Cuthbert, 2014], 483 and actually observable in catchments with non-uniform flow fields during long term 484 485 recession periods as in this field example. Hence such geometries do not affect the fundamental principle of deriving estimates of the R/S_v ratio by the method we have 486 proposed. Other deviations of field situations from the analytical model are also possible such 487 488 as differing drainage areas and streamflow timings for adjacent channels, lack of adjacent 489 channels, and non-parallel impermeable boundaries at differing distances. These should be considered on a case by case basis, and where significant deviations are found, modifications 490 to the methodology may be necessary to ensure accurate results. 491

492 **4.** Conclusions

We have developed a generalised conceptual model for understanding water table and groundwater head fluctuations in ephemeral stream catchments and, by accounting for the recession characteristics of a groundwater hydrograph, presented a simple but powerful new approach to quantifying indirect recharge both in the long term and on an event basis. Furthermore, a new, and globally unparalleled, data set of groundwater dynamics in a dryland

ephemeral stream catchment from Middle Creek, NSW, Australia has been used to test thetheoretical ideas developed using idealised models.

From examination of the extensive field dataset we find that head responses to ephemeral 500 streamflow events are controlled by pressure redistribution operating at three timescales from 501 vertical flow (days to weeks), transverse flow perpendicular to the stream (weeks to months) 502 and longitudinal flow parallel to the stream (years to decades). From application of the new 503 methods to the field dataset we find that, in relative terms, groundwater recharge increases 504 linearly away from the mountain front to the perennial stream section, and has a similar 505 spatial pattern both in the recent events analysed as well as over the longer term. In absolute 506 507 terms the long term indirect recharge estimates vary from approximately 30 to 80 mm/a with the main uncertainty in these values stemming from the challenge of being able to estimate 508 hydraulic properties at the appropriate spatial scale. 509

Further work will focus on the transferability of this approach to other dryland catchments 510 511 which have sufficient groundwater level data available. While we noted in the introduction 512 that multi-year observations of groundwater dynamics in ephemeral stream catchments are relatively rare, several data sets appear to show similar features to the data we have presented 513 here [Besbes et al., 1978; Carling et al., 2012; Goodrich et al., 2004; Hoffmann, 2007; 514 Houston, 2002]. Thus we expect that this methodology will be directly applicable to other 515 catchments. As longer time series become available from the Middle Creek catchment and 516 others that have recently been established in similar environments, for example as part of the 517 NCRIS groundwater infrastructure in Australia, the approach will be an important tool to 518 519 explore the relationship between groundwater recharge and climate change.

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Figure 1. Model of an idealised aquifer receiving indirect recharge from an ephemeral stream. The parameters used were as follows: $T = 200 \text{ m}^2/\text{d}$, $S_y = 0.01$. Dashed blue arrow represents the stream recharge boundary. Heads are relative to the fixed head boundary at x = 0 and represent the water table during a stream flow/recharge event. Numbers 1-4 are locations that represent the computed groundwater hydrographs in Fig. 2.



Figure 2. Output from the four locations in the model illustrated in Fig.1 showing superposition of transverse and longitudinal recessional characteristics. Grey shading indicates the period of steady flux input at the stream boundary. Black dashed lines show the exact proportionality between the variation of long term (straight line) groundwater head recession down the catchment and the long term recharge i.e. the long term recharge rate is equal to the specific yield multiplied by the long term head recession rate.



Figure 3. Rates of groundwater flux recession (D) after recharge ceases normalised to the recharge rate (R) used to determine the initial conditions, for variations in aquifer response time (ART = S_yL^2/T) and time, for a groundwater monitoring point positioned at x = 0.5L. Shaded zone is for $t < t_{lin}$ as defined by Cuthbert (2014).



Figure 4. Variation in the amplitude of the flux recession rate (A) normalised to the average recharge rate (R) plotted against the ratio of the ART (L^2S_y/T) to the period of recharge variation (P) and the position of the groundwater monitoring point with respect to the catchment boundaries (x/L). Plot created using Equation 8 from Cuthbert (2010). Away from the fixed head boundary, ART/P must be less than 1 for A/R to deviate significantly (more than 10%) from zero. i.e. variations in recharge at periods less than the ART will be damped out and not expressed as variations in the flux recession rate.



Figure 5. Groundwater hydrographs for the cases shown in Figure 2 with the long term
recession removed to reveal the effects solely due to recharge from the stream. Grey shading
indicates the period of steady flux input at the stream boundary.



Figure 6. Middle Creek and monitoring installations in the context of the Maules Creek
catchment. DEM used courtesy of Geoscience Australia. BF = Bellevue Farm; EC = Elfin
Crossing; MCF = Middle Creek Farm; EL = East Lynne.



Figure 7. Groundwater hydrographs (daily average), stream stage and cumulative rainfall. 672 Heads are given on the same vertical scale, but with the absolute values shifted to enable the 673 hydrographs to be compared. Dashed black arrows indicate stream flow events at East Lynne 674 (BH20-21) captured by an automatic camera, prior to the installation of stream stage 675 monitoring. Bold dashed black lines indicate periods of straight line groundwater recession. 676 The different piezometer screens for each borehole are coloured according to the key shown 677 678 with 1-4 being shallow to deep respectively. The time of slug testing is also marked as it led 679 to a temporary disturbance of the natural heads.



Figure 8. Groundwater hydrographs with background recessions 'removed'. The final head value represents the increase due to recharge from ephemeral stream flow since Jan 13. Where continuous data were not available from data loggers (i.e. BH22, BH19 & BH21), the residual heads were calculated using dip-tape measurements taken just prior to the stream flow event in late Jan 2013.



Figure 9. Comparison of long term and event based indirect recharge estimates.



Figure 10. Variation of estimated recharge with distance upstream and comparison of
observed dry period heads (i.e. during a straight line recession period) with heads predicted
using Equation 4.

	Recession	Residual	Error in	Long term	Recharge in	Error in
	rate	head	estimated	recharge	2013	recharge
		increase for	value of R/S_y			
		2013				
BH17	2678	2123	104	32	25	5.9
BH18	3118	2609	154	37	31	7.3
BH19	3166	2398	175	38	29	6.7
BH20	5711	4757	75	69	57	13.2
BH21	6444	5618	96	77	67	15.6
BH22	5892	4258	64	71	51	11.8

Table 1 Summary of recession rates and recharge estimates - all values in mm/a