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A hydrological framework for persistent river pools

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Abstract

Persistent surface water pools along non-perennial rivers represent an important water resource for plants, animals, and humans. While ecological studies of these features are not uncommon, these are rarely accompanied by a rigorous examination of the hydrological and hydrogeological characteristics that create or support the pools. Here we present an overarching framework for understanding the hydrology of persistent pools. We identified perched water, alluvial through flow and groundwater discharge as mechanisms that control the persistence of pools along river channels. Groundwater discharge is further categorized into that controlled by a geological contact or barrier (not previously described in the literature), and discharge controlled by topography. Emphasis is put on clearly defining through-flow pools and the different drivers of groundwater discharge, as this is lacking in the literature. A suite of diagnostic tools (including geological mapping, hydraulic data and hydrochemical surveys) is generally required to identify the mechanism(s) supporting persistent pools. Water fluxes to pools supported by through-flow alluvial and bedrock aquifers can vary seasonally and resolving these inputs is generally non-trivial. This framework allows the evaluation of the susceptibility of persistent pools along river channels to changes in climate or groundwater withdrawals. Finally, we present three case studies from the Hamersley Basin of north-western Australia to demonstrate how the available diagnostic tools can be applied within the proposed framework.

1 Introduction

Permanent or almost permanent water features along non-perennial rivers (hereafter referred to as "persistent pools") represent an important water resource for plants, animals, and humans. These persistent pools typically hold residual water from periodic surface flows, but also may receive input from underlying aquifers, and have alternately been termed pools (Bogan and Lytle, 2011; Jaeger and Olden, 2011; John, 1964), springs (Cushing and Wolf, 1984), waterholes (Arthington et al., 2005; Bunn et al., 2006; Davis et al., 2002; Hamilton et al., 2005; Knighton and Nanson, 2000; Rayner et al., 2009), and wetlands (Ashley et al., 2002). Non-perennial streams are globally distributed across all climate types (Shanafield et al., 2021; Messager et al., 2021). The occurrence of persistent pools along non-perennial streams has been well-documented (Bonada et al., 2020), particularly in the arid southwest of the U.S. (Bogan and Lytle, 2011) and across Australia (Arthington et al., 2005; Bunn et al., 2006; Davis et al., 2002). Several studies have confirmed that these water features support a highly diverse community of flora and fauna (Shepard, 1993; Bonada et al., 2020) and can vary significantly in water quality (Stanley et al., 1997). Persistent pools are also often of cultural significance (Finn and Jackson, 2011; Yu, 2000), providing key connectivity across landscapes for biota (Sheldon et al., 2010; Goodrich et al., 2018), and early hominid migration (Cuthbert et al., 2017). Paradoxically, the unique ecosystems they support are also sensitive to changing climate and human activities (Bunn et al., 2006; Jaeger and Olden, 2011). Persistent pools may dry out naturally after successive dry years (Shanafield et al., 2021) and recent studies have shown that persistent pools are also changing over time in response to alterations in climate and sediment transport (Pearson et al., 2020, Bishop-Taylor et al., 2017). However, their hydrology is typically poorly understood, and the treatment of the hydrology of persistent river pools in published literature to date has been largely descriptive, vague, or tangential to the main theme of the paper (Thoms and Sheldon, 2000). As a result, effective water resource management is limited by a lack of understanding of the mechanisms and water sources that support these persistent pools.

By far, the published literature on persistent pools focuses on the ecological processes and patterns. They have received attention for the role they play as a seasonal refuge (Goodrich et al., 2018), and with regards to connectivity between riparian ecosystems (Godsey and Kirchner, 2014). For example, they have been shown to host unique fish assemblages (Arthington et al., 2005; Labbe and Fausch, 2000), macroinvertebrate communities (Bogan and Lytle, 2011), and primary productivity (Cushing and Wolf, 1984). Recently, it was shown that the structure, but not composition, of these pools mirrors that of perennial rivers (Kelso and Entrekin, 2018). However, rarely are these ecological studies accompanied by a rigorous examination of the hydrological and hydrogeological characteristics that provide a setting for these ecologic communities. Although there are isolated studies that examine the composition of water and propose sources within specific pools (Hamilton et al.,2005; Fellman et al., 2011), more frequently they simply describe the seasonal persistence of flow and basic hydrologic parameters (typically temperature and salinity, sometimes also oxygen).

From a geological perspective, classification of persistent pools, and springs in general, dates back to the early 20th Century, when geological drivers such as faults and interfaces between bedrock and the overlying alluvial sediments were first discussed in relation to springs (Bryan, 1919; Meinzer, 1927). Subsequently, a diverse, modern toolbox of hydrologic and hydrogeologic field and analysis methods to analyse water source, age, and composition has evolved. Yet contemporary work on springs (Alfaro and Wallace, 1994; Kresic, 2010), and hydrogeology textbooks (e.g. Fetter, 2001; Poeter et al., 2020) are still based primarily on these early classifications. More recent classifications, moreover, are either descriptive or focus on the context (karst vs desert) or observable spring water quality (Springer and Stevens, 2009; Shepard, 1993; Alfaro and Wallace, 1994) and are not readily applied to understand the hydrology of persistent river pools (not all persistent pools are springs). There has also been a robust body of literature developed around surface water – groundwater interaction of the past 20 years (e.g. Stonedahl et al., 2010; Winter et al., 1998), some of which informs our understanding of persistent river pools, but has not yet been explicitly applied in this context. Similarly, our understanding of the hydrology of non-perennial streams and their links to groundwater systems continues to expand (Costigan et al., 2015; Gutiérrez-Jurado et al., 2019; Blackburn et al., 2021; Bourke et al., *In review*). Thus, there is both the need and opportunity for a comprehensive hydrologic framework (Costigan et

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al. 2016; Leibowitz et al., 2018) that incorporates the relevant literature on groundwater springs and surface - groundwater interaction, along with the modern suite of diagnostic tools, to provide a robust framework for understanding the hydraulic mechanism that support persistent river pools.

Here, we establish the conceptual models and nomenclature required for a more rigorous approach to the study of persistent river pools. We first classify the hydraulic mechanisms that support persistent pools (Section 2) and then critique the hydrologic tools available for identifying these mechanisms based on field observation (Section 3). We then discuss the susceptibility of persistent pools to shifts in climate or groundwater withdrawals based on the mechanism(s) supporting them (Section 4). Finally, we present three case studies from the Hamersley Basin of north-western Australia to demonstrate how the available diagnostic tools can be applied within the proposed framework (Section 5). In conclusion, we suggest next steps for refining and applying this framework to improve our understanding and management of persistent river pools (Section 6).

2 Hydraulic mechanisms supporting the persistence of in-stream pools

Here we propose a framework for classifying the key hydraulic mechanisms that support the persistence of pools along non-perennial rivers in environments where the shallow, unconfined aquifer does not support year-round flow (summarized in Table 1). Geologically, we start by considering the general case of a non-perennial river along an alluvial channel (inundated and/or flowing during contemporary flood events) within valley-fill sediments deposited over bedrock (Sections 2.1 and 2.2). We then move onto a discussion of the ways in which geological structures and outcrops can underpin the persistence of river pools by facilitating the outflow of regional groundwater (Section 2.3). The range of geological settings for non-perennial streams is vast (Shanafield et al., 2021); we have endeavoured to provide sufficient general guidance so that the principles can be applied to specific river systems as required. Hydrologically, we only consider the water balance of pools after surface flows have ceased and consider any water that has infiltrated to the subsurface saturated zone (which may be a perched aquifer) to be groundwater, irrespective of the residence time of that water in the subsurface.

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Identification of the hydraulic mechanisms supporting in-stream pools is essential for effective management of risks to pool ecosystems associated with groundwater withdrawals, changes to the hydraulic properties of the catchment (e.g. land use change) or climate change. The water balance of persistent pools may respond to a combination of more than one of these hydraulic mechanisms, and the dominant mechanisms can vary spatially and temporally within pools. For example, a pool may contain a mixture of water from streambed sediments and regional groundwater during certain hydroperiods, but the pool wouldn't persist through the dry season in that location without groundwater discharge from the regional aquifer. Thus, the maintenance of the stream ecosystem in its current state would require preservation of in-stream water storage and regional groundwater inflows.

2.1 Perched surface water

Perched surface water can be retained in topographic lows that retain rainfall and runoff during the dry season but are disconnected from the groundwater system (Fig. 1) if there is a low-permeability layer between the pool and the water table (Brunner et al., 2009). The presence of this low-permeability layer is essential to maintain a surface water body that is disconnected from the groundwater system. In the absence of a low-permeability layer, the surface water will slowly infiltrate into the subsurface (Shanafield et al., 2021). This low-permeability layer typically consists of clay, cemented sediments (e.g. calcrete) or bedrock (Melly et al., 2017). The persistence of water in these pools will depend on a) shading from direct sunlight and/or, b) sufficient water volume so that it is not completely depleted by evapotranspiration during the dry season (which will be a function of pool depth).

The occurrence and biological significance of such perched pools has been described particularly for rivers in inland Australia, where contribution of groundwater has been ruled out on the basis of pool hydrochemistry (e.g. Bunn et al.., 2006, Fellman et al., 2016). For example, along Cooper Creek in central Australia, geochemical and isotopic studies revealed a lack of connection to groundwater, and that convergence of flows at the surface and subsequent evaporative water loss-controlled water volumes in many pools (Knighton and Nanson, 1994; Hamilton et al., 2005). These pools are situated

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in depressions caused by erosion through sandy subsurface layers (note that the low-conductivity layer for perching was not elucidated). It should be noted, that definitive characterization of perched surface water (i.e. disconnected from the groundwater system) requires the measurement of a vertical hydraulic gradient between the water level in the pool and local groundwater, as well as identification of a low-permeability layer at the base of the surface water (Brunner et al. 2009). Although the ecological significance of perched in-stream pools is documented within the literature (Boulton et al., 2003; Arthington et al., 2005; Bonada et al, 2020), there is typically no detailed analysis of the hydrology and sampling is synoptic, so the mechanism of persistence is unclear.

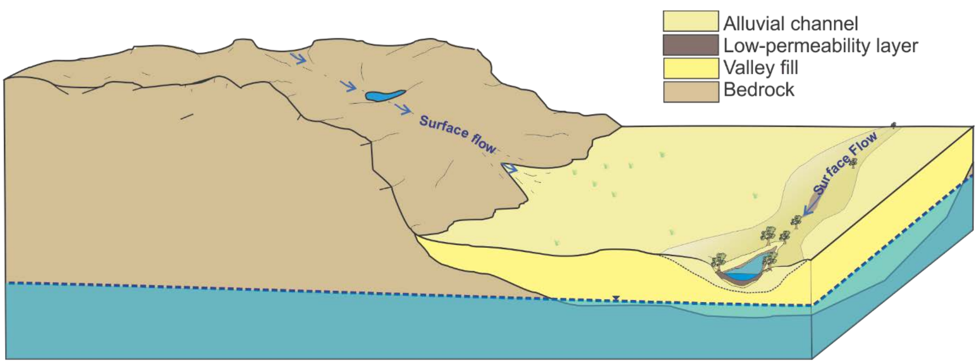

Figure 1 Schematic illustration of perched pools where rainfall-runoff collects in a depression that has morphology that limits evaporation and/or low permeability lithology beneath the pool that limit infiltration, allowing water to be retained for an extended duration.

2.2 Through-flow of alluvial groundwater

After a rainfall event, increases in water levels in rivers result in water storage and flow within the unconsolidated alluvial sediments in the beds and banks of stream channels (Cranswick and Cook., 2015). As the streamflow recedes after a flood, continuous surface flow ceases, resulting in isolated pools along the river channel. Water will remain within the alluvial sediments that line the stream channel beyond the period of surface flow, for a duration that will vary according to the amount of water stored, the hydraulic gradient within the sediments (from the headwaters to the catchment outlet) and the permeability of the sediments (Doble et al., 2012; McCallum and Shanafield, 2016). This alluvial water can be either perched above, or connected to, the regional unconfined aquifer depending on the depth of the regional water table and the presence of a low- permeability layer to enable perching (Villeneuve et al. 2015, Rhodes et al., 2017). Once within the alluvial sediments, this water can subsequently 1) flow through the alluvial sediments towards the bottom of the catchment 2) be lost to the atmosphere through evapotranspiration, or 3) migrate vertically downward into lower geological layers (Shanafield et al., 2021). Typically, a combination of these three processes occurs, and persistent surface water pools can be expressions of this water within streambed sediments (Fig. 2). Indeed, this source of water, limited to the floodplain, distinguishes the through-flow mechanism from regional groundwater discharge. The water level in these pools is effectively a window into the water table within the streambed sediments (Townley and Trefry, 2000).

The subsurface water flow through these disconnected pools can be hydrologically considered as an elongated, through-flow lake with inflow from the subsurface at the top of the pool and outflow to the subsurface at the bottom of the pool (Townley and Trefry, 2000; Zlotnik et al., 2009). The rate of inflow to (and outflow from) the pool is dependent on the hydraulic conductivity of the sediments (Käser et al., 2009) and the balance of inflow and outflow controls the depth and residence time of water in the pools (Cardenas and Wilson, 2007). The duration of persistence of the pool will also depend on the storage capacity of the alluvial sediments that support it; these pools may dry seasonally (Rau et al. 2017) or persist throughout the dry season if the water level in the alluvial sediments remains above the elevation of the pool. The water level and hydraulic gradients adjacent to persistent through-flow pools can change seasonally in response to alluvial recharge by rainfall events and subsequent depletion of water stored in the sediments. This process is analogous to “bank storage” adjacent to flowing streams (e.g. Käser et al., 2009; McCallum and Shanafield, 2016).

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There is a comprehensive body of literature on the dynamics of through-flow lakes (Pidwirny et al., 2006; Zlotnik et al., 2009; Ong et al., 2010; Befus et al., 2012). The storage and movement of water within alluvial sediments beneath and adjacent to streams has also been described extensively in literature on hyporheic exchange (e.g. Stonedahl 2010) with water fluxes across temporal (days to weeks) and spatial scales (centimetres to tens of metres). From a hydrological perspective, the key feature of the hyporheic zone, and hyporheic exchange, is that it is a zone of mixing between surface water and groundwater. Based on this definition, alluvial water that is perched above, and not connected to, the regional aquifer, does not fit the dominant conceptualization of hyporheic exchange. However, some authors have considered alluvial flows through this hyporheic lens (Rau et al., 2017, del Vecchia et al., under review) and the physical process that links streambed elevation changes to flow paths beneath pool-riffle sequences can be relevant to persistent in-stream pools, regardless of connection status.

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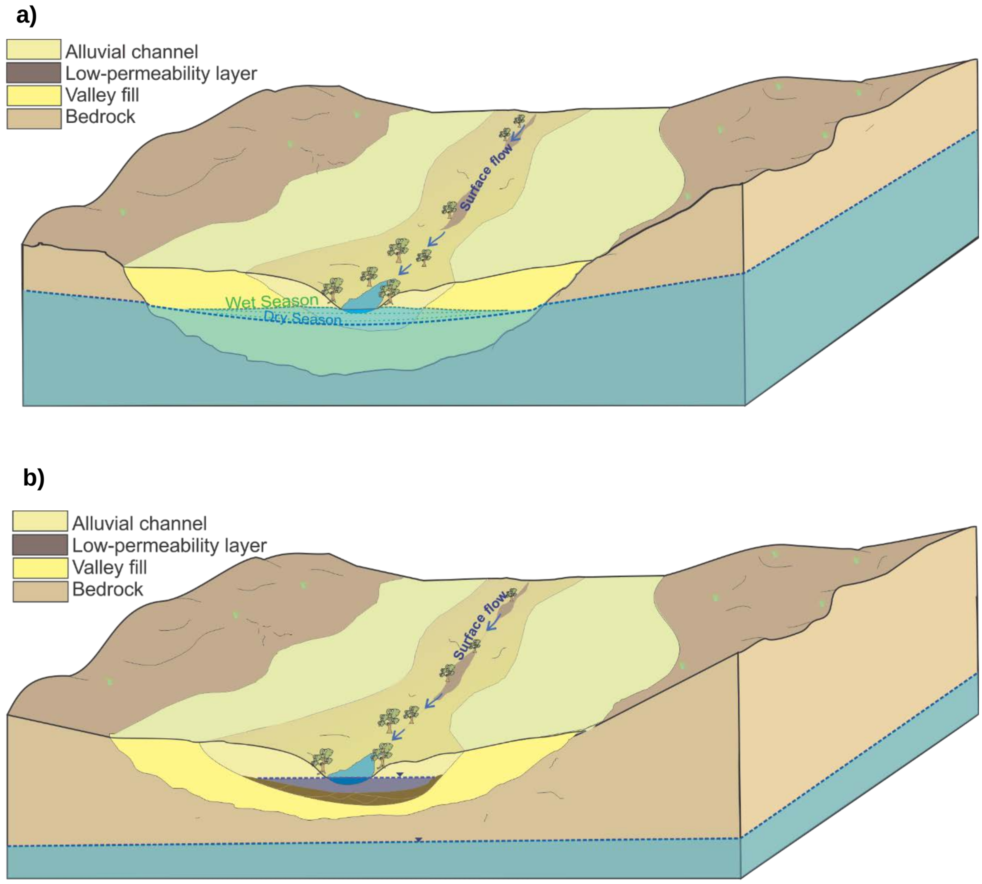

Figure 2 Schematic illustration of pools that are maintained by through-flow from the adjacent alluvial sediments. The water in these alluvial sediments can be either a) connected to the unconfined aquifer, or b) form a perched aquifer if the water is stored over a low-permeability geological layer.

2.3 Regional groundwater discharge

Similar to springs, rivers can be discharge points for regional groundwater, and this discharge can support the persistence of in-stream pools during periods without surface flow. Groundwater discharge through springs has been articulated into a range of detailed and complex categories, which are not consistent within the literature (Bryan, 1919; Springer and Stevens, 2009; Kresic and Stevanovic, 2010).

These existing spring classifications are based on geological mechanism, hydrochemical properties, landscape setting, or a combination of all three, leading to broad categories such as thermal or artesian, as well as nuanced distinctions based on detailed geological structures (Alfaro 1994). For the purposes of understanding persistent river pools, this array of categories is both overly complex and incomplete from a hydraulic point of view. For example, Springer (2009) presents a classification of springs based on their "sphere of influence", which is the setting into which the groundwater flows. A "limnocrene spring" is simply any groundwater that discharges to a pool, as distinct from say a "cave spring", which emerges into a cave. On this basis, one might consider all persistent pools that are not perched as limnocrene springs. However, the schema also articulates 'helocrene springs" which are associated with wetlands and "rheocrene springs" that emerge into stream channels. These also seem to be potentially fitting labels for persistent river pools, which does one choose? And what would it matter for water resource management and the conservation of pool ecosystems if you chose one category over the other?

We suggest two broad categories can encompass the range of hydraulic mechanisms supporting persistent pools in intermittent stream channels; geological features (i.e. lithologic contacts and barriers to flow), and topographic lows. This distinction is valuable because it facilitates an understanding of the source of groundwater discharge (shallow, near-water table vs deeper groundwater) and the size of the reservoir supporting the pool, both of which contribute to the susceptibility of pool persistence to groundwater pumping. This distinction can also be useful for identifying the dominant hydrogeological control on the influx of regional groundwater to the pool; in hard-rock settings with geological contacts and barriers the influx may be limited by fracture aperture, whereas in a topographic low the influx will be controlled by hydraulic head gradient between the pool and the groundwater source (see Case Studies below).

2.3.1 Geological contacts and barriers to flow

Geological contacts are well-established as potential drivers of groundwater discharge through springs (Bryan, 1919; Meinzer, 1927). For example, contact springs occur where groundwater discharges over a low-permeability layer, commonly associated with springs along the side of a hill or mountain (Kresic

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and Stevanovic, 2010; Bryan, 1919). Similarly, pool peristence can be supported by groundwater discharge into a stream channel over a low-permeability geological layer caused by the reduced the vertical span of the aquifer (Fig. 3a); where this vertical span reduces to zero is known colloquially as the aquifer "pinching out". This mechanism has been identified as driving regional groundwater discharge to streams (Gardener et al., 2011), but to our knowledge has not yet been explicitly discussed in the context of persistent river pools.

Outflow of groundwater where a catchment is constrained by hard-rock ridges that constrict groundwater flow (by reducing the lateral span of surface flow and the aquifer) can also support the persistence of surface water pools (Fig. 3b). Although the importance of catchment constriction has been identified by practitioners (e.g. Queensland Government, 2015), to our knowledge the discharge of groundwater caused by catchment constriction as a mechanism for surface water generation has not previously been described in published literature (springs or otherwise).

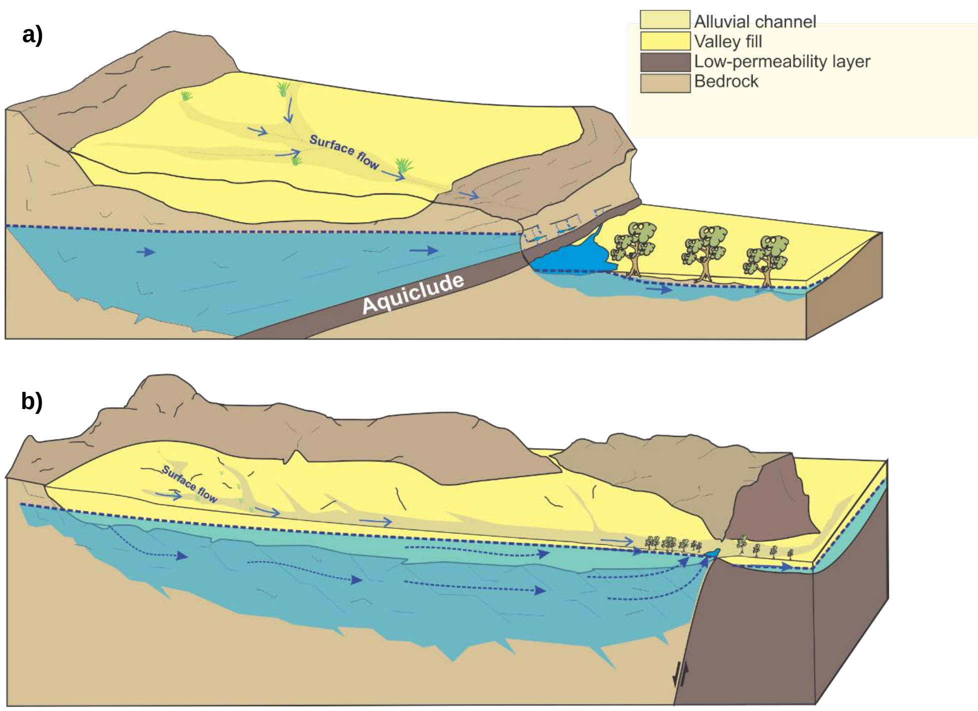

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Figure 3 Schematic illustration of a groundwater discharge pools where surface water persistence is driven by geological barriers that a) cause a regional aquifer to pinch out vertically, or b) form a lateral constraint on the catchment and underlying regional aquifer.

2.3.2 Topographically controlled seepage from regional aquifers

Pool persistence can be sustained by groundwater seepage from regional aquifers in the absence of geological barriers or contacts if there is a topographic low that intersects the regional water table (Fig. 4). This mechanism will generally occur where differential erosion causes a difference in topography, which is equivalent to depression springs (Kresic and Stevanovic, 2010; Bryan, 1919) and analogous to the lakes that form in pit voids left after mining ceases (McJannet et al., 2017). For example, pools likely supported by this mechanism have been identified within the Adelaide region of South Australia where erosion within a syncline has exposed bedrock, facillitating groundwater discharge (Lamontagne et al., 2021). Within the humid landscape of south-eastern USA, Deemy and Rasmussen (2017) also describe a vast number of pools along intermittent streams. These pools, which are seasonally connected by surface flows during the wet season, are expressions of the karst groundwater networks that underlie them and may be considered special cases of topographically-controlled groundwater discharge pools. Topographic depressions that fill seasonally with water, known as "sloughs" on the North American prairie, operate similarly hydraulically (seasonal snow melt inputs, evaporation induces groundwater inflow), but these sloughs are not within river channels and commonly reside within low-permeability glacial clays so that they are supported by the local-scale the groundwater system (Van der Kamp and Hayashi, 2009). Even some Arctic lakes, formed in shallow topographic depressions, receiving groundwater input and seasonally situated within a stream of snowmelt runoff (Gibson, 2002) can be considered as pools supported by topographically-controlled groundwater discharge.

Pools may also be sustained by topographically controlled seepage from confined aquifers if there is a fault or fissure that acts as a conduit to groundwater flow (different to Fig. 3a because there is no geological transition to sustain a hydraulic gradient across the pool). Topographically controlled

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discharge from a confined aquifer is analogous to artesian mound springs like those found in the Great Artesian Basin of central Australia (Ponder, 1986), but these do not reside within non-perennial streams. Groundwater discharge along fractures or faults has been identified as an important mechanism for groundwater discharge to the Fitzroy River in northern Australia (Harrington et al., 2013), but the significance of this regional groundwater discharge to individual persistent pools is not yet known.

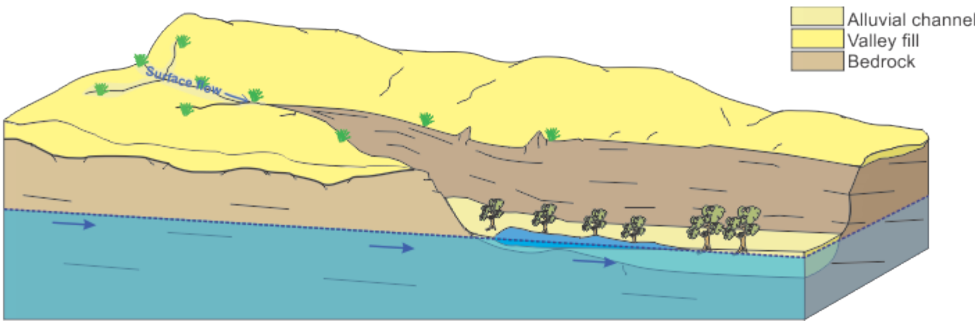

Figure 4 Schematic illustration of a pool receiving topographically-controlled groundwater outflow from an unconfined regional aquifer.

3 Diagnostic tools for elucidating hydraulic mechanisms supporting pool persistence

Several tools in the hydrologist's toolbox are appropriate for gathering the data needed to distinguish between the types of pools outlined in the previous section. For most of these, there are no examples specific to persistent pools along intermittent rivers. Therefore, in this section, general background and suggested considerations for use within persistent pools is given for a selection of the most common methods. The information these methods provide is critical to calculate water balances and identify susceptibility to groundwater withdrawals and climate change (Section 5).

The process of understanding pool occurrence is an iterative one. Data must be collected to infer the mechanism supporting the pool (e.g. geological mapping, water levels, salinity), but also an understanding of the pool mode of occurrence can be used to inform appropriate monitoring regimes. For example, pools that are supported by the discharge of deep regional groundwater are potentially vulnerable to groundwater abstraction, while perched pools are unlikely to be impacted. Thus, if managing impacts from groundwater abstraction, then monitoring efforts would be best directed to the groundwater-dependant pools at the expense of pools that are disconnected from the groundwater system. It is also important to note the potential logistical constraints that can apply when installing any infrastructure for sampling and monitoring in-stream pools. Persistent pools in arid landscapes are commonly sites of environmental and cultural significance (Finn and Jackson, 2011; Yu, 2000) so that appropriate approvals and permissions typically must be obtained prior to the installation of monitoring infrastructure. This may restrict the types of data that can be collected. Moreover, some sites may be sacred sites, limiting who is able to access them. Surface water features in general are a draw for travellers and roaming livestock, so that any infrastructure must be secure from theft or damage. Flood events and sudden, flashy streamflows are also potential threats to infrastructure, with substantial sediment and vegetation (branches, trees) transported across the floodplain to heights of 2-3 m that can (and have) destroyed sampling equipment. Furthermore, because regional groundwater inputs can be a relatively small (but important) component of the water balance of pools, snapshot sampling commonly targets the end of the dry season. This is when the contribution of regional groundwater is likely to be at its greatest. However, when un-seasonal or early rainfall occurs, or if infrastructure has been damaged, that endpoint in the water balance may not be captured.

3.1 Landscape position and remote sensing

Landscape position can provide some clues as to the mechanism controlling the persistence of a given pool. For example, a pool located high in the catchment on impermeable basement rock is likely to be a perched pool. A pool that is immediately prior to a ridge that constrains the catchment is likely to be supported by geologically constrained groundwater discharge. Lateral catchment constriction can commonly be identified from publicly available aerial imagery, but identification of vertical catchment constriction will usually require geological data from drilling or regional-scale geophysical surveys.

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The presence of geological contacts can be evident from readily available maps of surface geology, but the hydraulic properties of geological contacts are not known a-priori. Geological transitions can be zones of high permeability or barriers, or a combination of both (e.g. faults with high permeability in the vertical, low permeability laterally) depending on the depositional and deformational history of the area (Bense et al., 2013). Hydraulic head gradients can provide valuable insights, with a step-change in hydraulic head a key indicator for the presence of a hydraulic barrier. The presence of active deposition of geological precipitates can also be indicative of pool mode of occurrence with carbonates associated with groundwater discharge and subsequent degassing of CO_2 (Mather et al., 2019). Mapping the persistence of vegetation and water in the landscape based on remotely sensed data (i.e. NDVI or NDWI) can be used to identify pools that persist (Haas et al., 2009; Soti et al., 2009; Alaibakhsh et al., 2017), but this alone does not explain the hydraulic mechanism determining the location of the pool. Combining these vegetation indices with aerial geophysics (i.e. AEM) can aid in developing a better understanding of hydraulic mechanisms in remote areas, allowing the identification of low-permeability layers or geological structures that are not obvious from aerial photographs (Bourke et al., *In Review*).

3.2 Hydrography and pool water balances

Direct measurement of water balances in arid and semi-arid regions can be logistically difficult (Villeneuve et al., 2015). Rainfall (and therefore runoff) in arid and semi-arid environments is commonly patchy and water fluxes can be either too large to measure (streamflow during a cyclone) or too small to measure directly (dry-season groundwater seepage fluxes) (Shannon et al., 2002; Shanafield and Cook, 2014). In the absence of data to characterize pool hydrology, regional groundwater mapping can provide insights into the mechanisms supporting persistent pools, particularly if the geology has also been well-characterized (see Case Studies below for examples). Water table maps can articulate areas of groundwater recharge and discharge, and steep hydraulic gradients that may (but not definitely) reflect the presence of geological barriers (e.g. Fitts, 2013). For the ecologist,

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it is important to understand that regional-scale groundwater maps are always based on point-data of hydraulic heads measured in the groundwater system, interpreted by a hydrogeologist in the context of what is known about geology and surface drainage (Siegel, 2008). These maps can be refined based on measures of groundwater salinity and groundwater residence times (from environmental tracer data), both of which generally increase along a groundwater flow path. As such, these maps are limited by the spatial distribution of the data available (commonly sparse) and therefore may not accurately capture local-scale features and processes relevant to a particular pool of interest. Nevertheless, if an interpreted water table surface suggests that the regional water table is tens of meters below ground in the vicinity of a pool, then the surface water is likely (but not definitely) perched. If a pool is situated in a region that has been identified as a regional groundwater discharge zone, then this groundwater discharge is likely to be supporting pool persistence.

If instrumentation can be installed in the pool, then it may be possible to characterize the pool water balance. Once a pool becomes isolated from the flowing river, and in the absence of rainfall, a general pool water balance is given by;

$$\frac{\partial V}{\partial t} = Q_{\mathrm{i}} - \mathrm{Q_o} - EA \qquad (1)$$

where V is the volume of water in the pool (L^3), t is time (T), Q_i is the water flux from the subsurface into the pool (L^3T^{-1}), Q_o is the water flux out of the pool into the subsurface (L^3T^{-1}), E is the evaporation rate ($L\ T^{-1}$) and A is the surface area of the pool (L^2). The water level in the pool, h_{p} (L), can be routinely measured by installing pressure transducers, but conversion of water levels to pool water volume requires knowledge of pool bathymetry, and the relationship between h_{p} and V will change during the dry season as the pool water level recedes, or if pool bathymetry is altered by scour and/or sediment deposition during flood events. Evaporation rates can be taken from regional data or empirical equations, but actual losses can vary depending on solar shading, wind exposure and transpiration (McMahon et al., 2016). For pools with visible surface inflow or outflow, these rates can potentially be measured using flow gauging (or dilution gauging), but relatively small flow rates and bifurcation of flow can make this challenging.

Modified versions of this general water balance can be defined for particular pools, depending on the hydraulic mechanism(s) supporting pool persistence (Table 1). For perched pools, which are disconnected from the groundwater system, Q_i=$Q_o = 0$, so that the only component of the water balance is water loss through evaporation. Pools that are supported by alluvial through-flow are hydraulically connected to the water stored in the streambed alluvium. Water levels within this alluvium will be more dynamic than regional groundwater levels, so that influx and efflux rates that can change over time in response to rainfall events or seasonal drying (of the near-subsurface). For pools supported by groundwater discharge, influx will dominate over efflux ($Q_i > \mathrm{Q_o}$). If the groundwater discharge is over an impermeable aquiclude (see Fig. 3b) there will commonly be a seepage zone up-gradient of the pool so that water influx is via surface inflow, but outflow to the subsurface can form a source of groundwater recharge to the adjacent (down-gradient) aquifer. If the groundwater discharge is controlled by topography, then the pool will be a site of regional groundwater discharge so that local groundwater recharge (and Q_o) should be negligible.

If a pool is connected to the groundwater system Q_i (or Q_o) can be estimated from Darcy's Law;

$$Q_\mathrm{i} = K\frac{\Delta h}{\Delta x}A_\mathrm{i} \qquad (2)$$

where K is hydraulic conductivity, $\frac{\Delta h}{\Delta x}$ is the hydraulic gradient between the pool and the source aquifer, and A_i is the area over which the groundwater inflow occurs (which will usually be less than the total area of the base of the pool). The major limitations of this approach are that K of natural sediments varies by ten orders of magnitude (Fetter, 2001), and that the area of groundwater inflow needs to be assumed or estimated using a secondary method. Hydraulic gradients between pools and streambed sediments can be measured using monitoring wells or temporary drive points, with Δh usually on the order of centimetres at most. Determination of the hydraulic gradient between regional aquifers requires that the water level in the pool has been surveyed to a common datum and there is a monitoring well near the pool to measure the groundwater level relative to that datum. In shallow, groundwater

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dominated lakes, geophysical methods have also been used to determine local hydraulic gradients, and therefore the direction of the water flux(es) between groundwater and surface water (Ong et al., 2010; Befus et al., 2012). Blackburn et al (2021) similarly applied shallow geophysical surveys,. combined with mapping of hydraulic conductivities, to identify they key structures and processes controlling water fluxes between groundwater systems and the streams that host persistent pools (Blackburn et al., 2021).

3.3 Tracer techniques and pool mass balance

Numerous studies of streams and lakes have employed hydrochemical and mass balance approaches to quantify water sources (Cook, 2013; Sharma and Kansal, 2013) and groundwater recharge (Scanlon et al., 2006). Some of these methods are also applicable in persistent pools, but may require modification, or an iterative approach that allows for refinement of the methods as the mechanism supporting the pool is elucidated. In its simplest form, snapshot measurements of pool hydrochemistry (salinity, pH, major ions) can help distinguish pools that are connected to groundwater from those that are not (Williams and Siebert, 1963). Dissolved ions and stable isotopes of water are relatively cheap and easy to measure and have been used extensively to estimate recharge/discharge, groundwater flow, and ecohydrology in arid climates (Herczeg and Leaney, 2011). However, their application to identify or quantify water sources can be limited by overlapping values (Bourke et al., 2015), and spatiotemporal variability (see Case Studies). Time series of electrical conductivity (EC) and stable isotopes through flood-recession cycles can indicate relative rates of evaporation and through-flow (Siebers et al., 2016; Fellman et al., 2011) and allow identification of the hydraulic mechanism(s) supporting pool persistence. For example, if a pool is supported by regional groundwater discharge, the re-equilibrated with the groundwater EC value during the dry-season (provided there isn't another streamflow event); in a perched pool, the pool EC will not plateau, but continue to evapo-concentrate until the next flood event. Stable isotopic values of pool water can be interpreted similarly; groundwater seepage from a regional aquifer will have a relatively consistent isotopic value, while a pool isolated from the groundwater source will experience isotopic enrichment through evaporation (Hamilton et al., 2005). Pools receiving alluvial throughflow will have isotopic values that reflect the balance of inputs (from alluvial groundwater) and outputs (evapotranspiration and outflow to alluvial groundwater). The isotopic values in the alluvial water itself can also become enriched through evapotranspiration during the dry season resulting in variability over time, and throughout the catchment, so that end-member values should be defined locally. In one case, strontium isotopes were found to be more useful than stable isotopes of water for identifying groundwater contributions to in-stream pools because the concentration in the groundwater end-member was far more constrained than salinity or stable isotope values (Bestland et al., 2017). Importantly, the interpretation of hydrochemical data should ideally be supported by a robust understanding of the surrounding geology to ensure that the hydraulic mechanisms identified are physically plausible. For example, Fellman et al. (2011) identified a number of perched pools along an semi-arid zone alluvial stream channel, but in the absence of a low-permeability layer within the alluvium (which was not identified) it is unclear how pool water would persist in the absence of hydraulic connection to alluvial groundwater (or regional groundwater discharge).

Radon-222 is a commonly applied tracer in studies of surface water – groundwater interaction, and ^{222}Rn mass balances have been effective for quantifying groundwater contributions to streams and lakes (Cook, 2013; Cook et al., 2008). Preliminary measurements of ^{222}Rn in persistent pools indicates substantial spatial variability in ^{222}Rn activity along the pools, reflecting the spatial distribution of groundwater influx and gas exchange. This spatial variability will limit quantification of groundwater discharge based on ^{222}Rn mass balance but can allow for hot-spots of groundwater discharge to be identified (see Case Studies). Other groundwater age indicators have been measured along streams to identify groundwater sources (Gardener et al., 2011; Bourke et al., 2014), but their applicability in pools is yet to be determined. Given that shallow, stagnant water is common, tracers such as ^{14}C or ^{3}H, which don't rapidly equilibrate with the atmosphere (Bourke et al., 2014; Cook and Dogramaci, 2019), are likely to be better than gaseous isotopic tracers (e.g. ^{4}He) that equilibrate rapidly (Gardner et al., 2011). If a mass balance approach is applied, then hydraulic measurements to constrain the pool water balance should be made in conjunction with hydrochemical sampling to ensure that the water balance is appropriately reflected in the mass balance.

Temperature measurements have been used extensively to identify and quantify water fluxes across streambeds and lakebeds (e.g. Shanafield et al.,2010; Lautz, 2012). Diel amplitudes of subsurface temperatures have been used to identify the transition from flowing stream to dry channel (with isolated pools) in ephemeral systems (Rau et al., 2017). In persistent pools, temperatures at the water sediment interface can be used to map zones of groundwater inflow (Conant, 2004). In arid zones, groundwater temperatures will often be warmer than pool temperatures and this type of survey is best conducted at dawn when the temperature gradient between pool and groundwater is at a maximum and there are no confounding effects from direct solar radiation. This mapping can be conducted using point sensors or thermal cameras, but in natural water bodies this method has primarily found success at thermal springs where the temperature difference between surface waters and groundwater inflows is on the order of 10 °C (Briggs et al., 2016; Cardenas et al., 2011). Vertical profiles of temperature can also be used to estimate vertical fluid fluxes but the application of this approach in pools with coarse alluvial sediments (commonly through-flow pools) is likely to be limited by lateral flow within the subsurface when $K_h>K_v$ (Rau et al., 2010; Lautz, 2010). Analytical solutions for temperature-based flux estimates also break-down at low flux rates where the difference between convection and conduction is difficult to determine (Stallman, 1965). Recently developed instrumentation for measuring 3D flux fields (Banks et al., 2018) shows promise, but installation in course alluvial sediments like those commonly found in arid streambeds remains a challenge. Point-scale measurements also require up-scaling and these methods may not be applicable in fractured hard-rock pools.

4 Management implications: Susceptibility of persistent pools to changing hydrological regimes

Robust water resource management in semi-arid regions requires an understanding of the ways in which human activities or shifting climates can alter water balances and/or the duration of pool water persistence (Caldwell et al., 2020; Huang et al., 2020). In the absence of published literature quantifying the susceptibility of persistent pools, we present general guidance on the susceptibility of pools to changes in rainfall and groundwater withdrawals based on hydrologic principles (Table 1). Intuitively, the size of the reservoir (surface catchment or groundwater storage) that supplies water to the pool should be a key factor in determining the susceptibility of persistent pools to changing hydrological regimes. However, the patchiness of rainfall and substantial transmission losses typical of semi-arid zone intermittent river catchments (Shanafield and Cook, 2014) mean that for pools reliant on surface catchments (perched or supported by alluvial through-flow), catchment size alone is unlikely to be a robust predictor of resilience. As has been demonstrated for arid zone wetlands in Australia (Roshier et al., 2001), pools that are storage-limited can be highly sensitive to climate variability. However, increasing heavy rainfall events my not necessarily result in increased pool persistence (particularly in pools closest to the location of rainfall) if subsurface storage up-gradient of the pool is already filling during the wet season. In this case, subsequent rainfall will increase streamflow downstream, but not result in increased subsurface storage in the reservoir supporting the pool. Moreover, recent work has shown that groundwater response times are sensitive to aridity, with longer response times associated with increased aridity (Cuthbert et al., 2019), so that there may be substantial time-lags between climate variability and hydrologic response in pools supported by groundwater discharge.

We have distinguished between geological or topographic control on groundwater discharge, but this distinction may not always be critical from a management perspective. In any system connected to groundwater, perturbation of the dynamic equilibrium between groundwater recharge and discharge can impact surface water-groundwater interactions; the timing and extent of the change will depend on the magnitude and rate of alteration (Winter et al., 1998). The hydraulic head gradients (and groundwater discharge rates) supporting persistent river pools may be small (Δh on the order of cms), so that small decreases in groundwater level (either due to successive low-rainfall years, or groundwater withdrawals) can potentially have a detrimental impact on the pool and cause the pool to dry out (particularly for topographically controlled groundwater discharge to pools). For pools supported by alluvial through-flow, the water balance (Table 1) is dominated by water outflow from contemporary fluvial deposits but abstraction from regional groundwater could impact the pool if these two subsurface

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reservoirs are hydraulically connected. The volume of groundwater storage in the source reservoir can indicate the resilience of pools to hydrological change (i.e. a longer groundwater system response time), but impacts will also depend on the distance from the recharge zone or groundwater abstraction (Cook et al., 2003). The time-lag prior to a decrease in groundwater outflow to the pool, and shape of the response (i.e. a slow decline or sharp decrease), will also depend on the spatial distribution of the forcing (pool distance from recharge or groundwater abstraction) (Cook et al., 2003; Manga, 1999). Thus, focussed groundwater abstraction close to a pool will cause a larger and faster reduction in groundwater outflow than diffuse abstraction across the aquifer, or abstraction further away (Cook et al., 2003; Theis, 1940). For example, groundwater pumped from within 1 km of a pool will result in a rapid decrease in discharge (months to years) but the same volume of abstraction distributed throughout the catchment will result in a more gradual decline in groundwater discharge to the pool (years to decades). Susceptibility can be further modified by geological barriers, which may not be obvious from the surface topography or regional geological maps (Bense et al., 2013), but can isolate pools from the regional groundwater system and either i) increase susceptibility to pumping within the connected aquifer, or ii) reduce susceptibility if the pumping is on the other side of the barrier (Marshall et al. 2019).

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Table 1 Summary of hydrological framework for persistent pools

Mechanism supporting pool persistence and water balance*	Physical characteristics	Hydrochemical characteristics	Susceptibility to stressors
Perched water $\frac{\partial V}{\partial t} = EA$	Topographic low that catches rainfall/runoff. Present in i) elevated hard-rock headwaters of catchments and ii) regionally low-lying topographic location. Water levels in aquifer lower than pool water levels. Vertical head gradient between pool and aquifer with unsaturated zone below pool.	Highly variable; hydrochemistry is a function of rainfall and subsequent evaporation. Substantial enrichment of solutes and water isotopes during dry season. Precipitated salts usually wash away in next flood, (or do not form because of low solute concentrations in streamflow source)	Relies on surface flows and overland runoff, which is directly tied to precipitation. Sensitive to climate but largely independent of groundwater use. Where infiltration capacity is high pools in downstream areas are more vulnerable to reduced rainfall.
Alluvium through-flow $\frac{\partial V}{\partial t} = Q_i - Q_o - EA$	Expression of river alluvium water table and through-flow. Head gradient reflects water table in alluvium. Water levels in pool coincident with water level in adjacent alluvium (cm-scale gradients expected at influent or effluent zones). Bank storage is important for pool water balance. Absence of surface geological features (e.g. hard-rock ridges) or waterfalls. Physical location may migrate as flood-scour re-shapes alluvium bedform.	Hydrochemically similar to alluvial water; enrichment of solutes and water isotopes during dry season limited by through-flow. Flood water flushes through the alluvium and replaces or mixes with any residual stored water (i.e. hydrochemically flood and alluvial groundwater are the same after a flood). More through-flow means shorter pool residence time and less enrichment.	Relatively small changes in rainfall or groundwater level can result in pool drying if the water level in the unconfined (alluvial) aquifer is reduced to below the base of the pool. Impact of withdrawals from alluvium depends on volume and proximity to pool. Abstraction from regional aquifers that are hydraulically connected to alluvium may also affect pool water levels by inducing downward leakage from alluvium.
Groundwater discharge			
1) Geological contacts and barriers to flow $\frac{\partial V}{\partial t} = Q_i - Q_o - EA$	Two sub-types: i) Catchment constriction across ridges, or ii) aquifer thinning due to geological barrier intersecting topography. Presence of waterfalls or surface geological features (hard-rock ridges). Hydraulic head step-changes across pool feature. Carbonate deposits if source aquifer has sufficient alkalinity.	Consistent hydrochemical composition at point of contact/barrier. Evapo-concentration and evaporative enrichment down-gradient of discharge point. Initial pulse of water from runoff may be saline, pool salinity equilibrates with groundwater at low water levels.	Susceptibility to groundwater abstraction depends on scale of source groundwater reservoir (if large then potentially more resilient) and location of groundwater abstraction. Water persistence is less susceptible to changes in rainfall than other pool types. Presence of geological barrier between pool and groundwater abstraction may limit impacts.
2) Topographically controlled seepage from regional aquifer $\frac{\partial V}{\partial t} = Q_i - EA$	Topography intersects i) water table or ii) preferential flow from artesian aquifer. Standing water persists during dry season due to groundwater discharge in absence of rainfall. Negligible recharge to aquifer during flood event (pool is regional discharge zone). Carbonate deposits if source aquifer has sufficient alkalinity.	Consistent hydrochemical composition at point of seepage. Initial pulse of water from runoff may be saline, pool salinity equilibrates with groundwater at low water levels	Susceptibility to groundwater abstraction depends on scale of source groundwater reservoir (if large then potentially more resilient) and location of groundwater abstraction. Hydraulic gradient supporting pools may be similar to pool depth. No geological barrier to limit susceptibility.

*Water balance of residual pool when disconnected from surface water flows and if only the one mechanism is operating

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5 Application of this framework to persistent pools in the Hamersley Basin

In this section we demonstrate the application of this framework to persistent river pools in north-west Australia. We begin by providing an overview of our understanding of the hydrology of persistent river-pools in the Hamersley Basin region. We then present three case studies to demonstrate how some of the tools described in Section 3 can be applied to identify the key hydraulic mechanisms supporting pool persistence, and the implications for pool susceptibility.

5.1 Overview of persistent pools in the Hamersley Basin

The Hamersley Basin has an arid-tropical climate with a wet season from October to April and a dry season from May to September (Sturman and Tapper, 1996). Average annual rainfall is less than 300 mm yr^{-1} with most rain falling between December and April (www.bom.gov.au). Annual rainfall statistics can vary dramatically, depending on the influence of thunderstorms and cyclone activity. Thunderstorm activity is commonly highly localised, limiting the potential for spatial interpolation of data from individual monitoring sites. Annual evaporation is around 3000 mm yr^{-1} (www.bom.gov.au), or about ten times annual rainfall, so that permanent surface water is rare. Ranges, spurs, and hills are separated by broad alluvial valleys with numerous deep gorges created by differential erosion. During large flood events, runoff creates sheet flow along the main channel and the extensive floodplain can remain flooded for several weeks. In the absence of cyclonic rainfall, surface water is generally limited to a series of disconnected pools along the main channels. The valleys are filled with up to 100 m of consolidated and unconsolidated Tertiary detrital material consisting of clays, gravels, and chemical precipitates. The Quaternary alluvial sediments along the creek-lines and incised channels (incised on the order of metres) consist of coarse, poorly sorted gravel and cobbles (thickness of up to tens of metres, widths of up to hundreds of metres). Fresh groundwater is abundant throughout the region, both within the Archean basement rocks, where permeability is increased via weathering, fracturing or mineralisation, and within the Tertiary and Quaternary sediments (Dogramaci et al., 2012).

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Numerous persistent water features have been identified along drainage lines that span the range of hydrogeological mechanisms in the framework outlined in Section 2 (Fig. 5). A sub-set of these (22 pools) have been investigated in more detail (Fig. 6) to characterize their mode of occurrence (Dogramaci, 2016). Based on data from this subset of pools, we have generalized the distribution of the hydrogeologic mechanisms supporting pool persistence across this landscape (Fig. 7). Perched pools are generally found in elevated, hard-rock areas where erosion has created a deep pool that is shaded to minimise evaporation. For example, there are approximately 20 pools that reside within the ephemeral drainage lines of the Western Range that flow for a few days in response to rainfall; a subset of pools that are deeply incised and shaded persist all year round (those that are shallower and more exposed to sunlight dry out faster and are not perennial). These pools are important ecologically (supporting bat populations) and culturally (supporting traditional hunting practices). Because these pools are not connected to groundwater, they are not directly at risk of depletion by groundwater withdrawals. However, they are susceptible to changes in streamflow that reduce the water storage in the pools at the commencement of the dry-season, either due to reduced inflows or in-filling by sedimentation.

Persistent pools that are connected to groundwater are also abundant across the Basin, with the folded and tilted layered sedimentary sequence resulting in numerous exposures of geological contacts at the land surface. Groundwater discharge from the unconfined aquifer through contact springs is therefore a common mechanism supporting persistent river pools in this region. These are particularly prevalent at the intersection of fluvial deposits and erosion-resistant, low permeability basement rocks. Groundwater-fed pools are also present due to catchment constraints where erosion-resistant layers form ridges in the landscape. Pools supported in part (or completely) by alluvial through-flow are also common along the stream channels due to the storage capacity of the coarse alluvial sediments. The hydraulic resistance caused by a catchment constraint can further enhance the persistence of alluvial water storage up-stream of the constraint, resulting in numerous persistent pools supported by alluvial throughflow. Although these pools are supported by groundwater, the hydraulic gradients maintaining

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groundwater inflow to the pool are commonly on the order of tens of centimetres, so that relatively small changes in the water balance can result in the pool drying out (e.g. successive low-rainfall years).

Figure 5 Photos of persistent pools within the Hamersley Basin, spanning the range of hydrogeological mechanisms within the proposed framework.

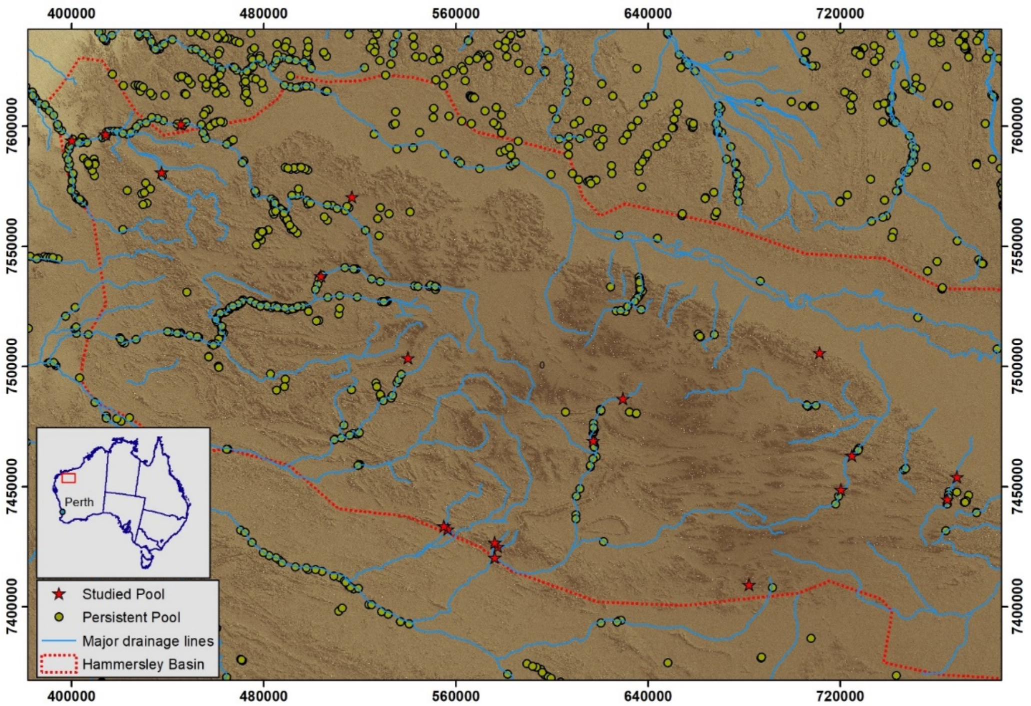

Figure 6 Prevalence of persistent pools on watercourses in the Hamersley Basin ("Waterholes" features from Geodata Topo 250K Series 3 data set, http://pid.geoscience.gov.au/dataset/ga/63999) and select pools examined in detail.

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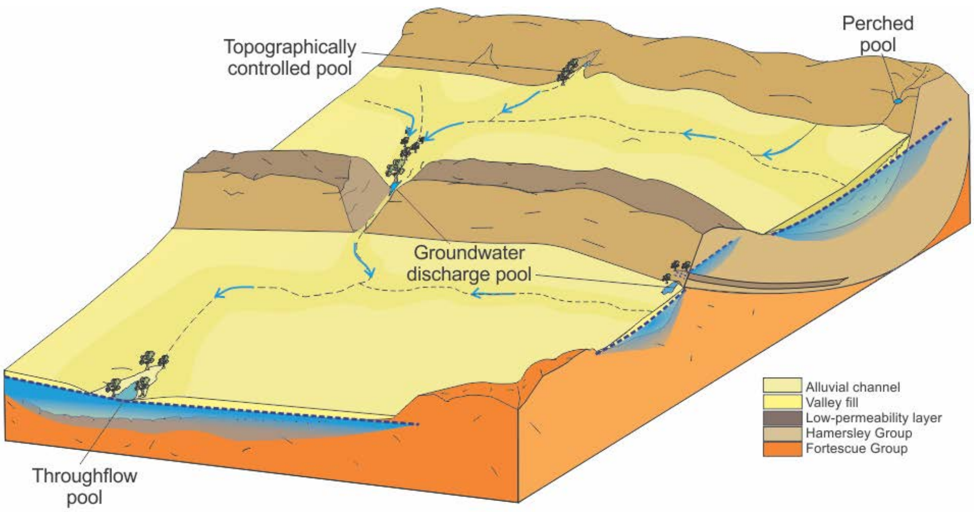

Figure 7 Generalized landscape position of each type of persistent pool within the Hamersley Basin.

5.2 Case Studies

The following three case studies demonstrate the application of this framework to three different pools (or pool systems) within the Hamersley Basin. To the best of our knowledge these pools have not been impacted by human activities. These case-studies demonstrate the application of key methods to infer hydraulic mechanisms supporting pool persistence, and the complexity of applying these methods in real-world situations. We start with a simple case, and build complexity with each case study, using data that highlight the temporal and spatial variability in pool hydrochemistry and provide valuable insight into the supporting hydraulic mechanisms (but also limits the appropriateness of basing an assessment on a small number of samples). The implications of these mechanisms for the susceptibility of the pools to groundwater withdrawals or changing climate are also discussed.

5.2.1 Case study 1: Plunge Pool

Plunge Pool (Fig. 8a) is located at the base of a steep topographic drop-off that exposes the Marra Mamba Formation (fractured banded iron formation, shale and chert). The Wittenoom Formation (consisting of dolomite and shale) and Marra Mamba Formation are hydraulically connected and form an unconfined regional aquifer where there has been sufficient weathering and fracturing to generate secondary porosity. This aquifer is 50-100 m thick and divided laterally by (sub-)vertical dykes on the order of 1 km apart (but as close as 100 m) that act as hydraulic barriers within the groundwater system. The surface catchment has an area of approximately 26 km^2 and is storage limited. Regional groundwater in the adjacent aquifer has a hydraulic head of 547 m AHD at a distance of 200 m from the pool, increasing to 557 m AHD 600 m from the pool, indicating the presence of a geological barrier between these two monitoring wells. Seasonal variation in groundwater hydraulic heads is minimal (on the order of 0.2 m).

The pool is perennial with seasonal water level fluctuation driven by variation in streamflow, groundwater inflow and evapotranspiration (Fig. 8b). The varying proportions of the pool water balance components are reflected in the temporal variation in the salinity of water in the pool. At the onset of the first wet season flood the salinity in the pool spikes (up to 4171 mS/cm), reflecting the flushing of surficial salts that were deposited during the previous dry through the catchment. Subsequent rainfall events then cause a rapid freshening of the pool (to as low as 124 mS/cm within 1 day). In the absence of rainfall, the salinity of the pool equilibrates to the that of groundwater in the regional aquifer (900 mS/cm). Given the consistency of groundwater levels, this inflow rate will be relatively constant, so that (in the absence of streamflow) the variability in the salinity of the pool is driven by seasonal variation in temperature and evapotranspiration (Bureau of Meteorology Station #007185, Paraburdoo Aero). These seasonal weather patterns drive evapo-concentration of solutes in the pool as water levels fall during the dry season and freshening of the pool as water levels rise when evapotranspiration decreases in winter (May-Sep).

Based on these data, the dominant hydraulic mechanism supporting the persistence of this pool is groundwater inflow from the regional aquifer that is intersected by topography (Section 2.3.2). In spite of the source being a regional aquifer, the spatial extent of the groundwater reservoir supporting the pool is limited by the presence of geological dykes (Fig. 8c). The pool effectively acts as a “drain” on the underlying/adjacent compartment of the unconfined aquifer with the inflow rate to the pool controlled by the hydraulic conductivity of the aquifer (variation in groundwater levels is negligible). The pool is also hydraulically connected to the alluvial aquifer and water from the pool is likely to infiltrate into the alluvium on the down-gradient side, but this has not been measured directly (alluvium is absent up-gradient of the pool – therefore alluvial through-flow is not a supporting mechanism). The susceptibility of this pool to groundwater withdrawals is controlled by the hydrogeological compartmentalization. The pool will be more susceptible to groundwater withdrawals from the aquifer between the dyke and the pool, and less susceptible groundwater withdrawals outside of this compartment. Given that evaporation is an important component of the water balance and contributes to the regulation of water levels, this pool is also susceptible to increases in evapotranspiration that are predicted as temperatures increase under climate change (IPCC, 2021).

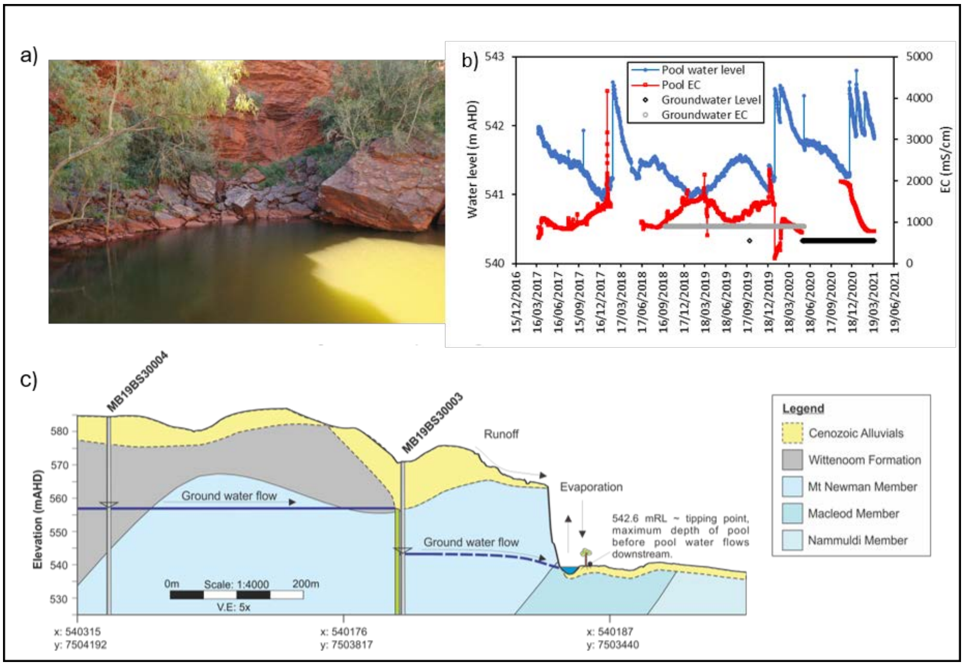

Figure 8 a) photo of Plunge Pool, b) pool water level and electrical conductivity (EC), c) hydrogeological setting of the pool.

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5.2.2 Case Study 2: Howie's Hole

Howie's Hole is a pool within stream channel alluvium at the exit point of a short, narrow gorge (Fig. 9a). Immediately at the outlet of the gorge (approximately 30 m up-hydraulic gradient of the pool) there is also a seep where groundwater outflows to surface for most of the year (seep dries for approximately 2-3 months at end of dry season). The seep is supported by the regional unconfined aquifer hosted within the Marra Mamba Formation (fractured BIF, shale and chert) and the surficial sediments above it (including the alluvial channel sediments), which are hydraulically connected. At the seep, the Brockman formation has become adjacent to the Marra Mamba due to faulting and this forms a relatively impermeable hydraulic barrier approximately 700 m wide (identified by the abrupt change in water table depth either side of the formation). The surface catchment upstream of Howie's Hole has an area of 33 km^2. The gorge restricts the stream channel from 30 m width down to a channel width of 10 m, enhancing the flow rate and resulting in scour and erosion of the Brockman formation. This area of scour during high-flow events has subsequently been filled by deposition of unconsolidated alluvial sediments, which are now at the base of the pool (sediments speculated to be 5-10 m deep).

The height of the regional water table is only known 1.5 km away from the seep, with seasonal fluctuations of 1-2 m (Fig. 9b). We assume that the water table declines towards the seep consistent with topographic elevation change (~ 20 m drop over 1.5 km), and that the seep reflects the height of the water table at that location (elevation of the groundwater seep is 405 m AHD). During the period of observation, the groundwater seep dried up when the measured water table elevation dropped below ~418.4 m AHD (water sample collected when the measured water table was at 418.5 m AHD on 12th Nov 2018); the seep was dry when the measured water table was at 418.3 m AHD on 7th Dec 2018. Pool water levels track groundwater elevations above 418 m AHD, but data from 2019 shows the pool depth levelling off as the water table at the monitoring bore drops below 418 m AHD, suggesting the cessation of significant groundwater inputs. The pool water levels have not

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been surveyed to the Australian Height Datum, but pool water level is consistently below the elevation of the seep (approximately 398 - 400 m AHD).

Similar to Plunge Pool, the pool salinity spikes with the seasonal onset of rainfall, before freshening once the accumulated salts have flushed through (Fig. 9c). In the absence of rainfall, pool salinity is similar to groundwater at the water table (Marra Mamba EC 1140 uS/cm) until the water table drops below the pool and groundwater inputs (from the seep) cease. Subsequently, evapo-concentration dominates the water balance of Howie's Hole, resulting in salinity increases. This process of disconnection from regional groundwater is also evident in stable isotopic values at the site (Fig. 9d). Isotopic values at the seep are relatively constant and close to values in the pool while the groundwater is connected; after disconnection (Aug 2018) isotopic values increase in response to evapo-concentration.

Based on these data we conclude that Howie's Hole reflects the water level in the alluvial aquifer within the stream channel (Fig. 9e). The location of the groundwater seep is determined by the geological contact between the permeable Marra Mamba Formation and impermeable Brockman Iron Formation in the subsurface, which coincides with the catchment constriction (gorge) that forms an outlet for surface and groundwater. As a result of the streamflow regime caused by this catchment constriction, the Brockman Iron Formation has been eroded and subsequently filled with unconsolidated stream channel sediments; water storage within these sediments now support the persistence of this pool.

The water level and isotopic data indicate a threshold groundwater level for inflow of groundwater to the pool, such that the pool water balance is primarily dominated by groundwater recharged during the previous wet season. Below this threshold water level for groundwater inflow, the persistence of the pool relies on local water storage within the streambed alluvium (supporting pool depths of up to 0.2 m). The persistence of this pool is therefore susceptible to 1) wet season rainfall that is inadequate to recharge the unconfined aquifer to above the threshold water level, or 2) groundwater withdrawals that reduce seasonal peak groundwater levels to below the threshold level.

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In the absence of this groundwater inflow, the pool is supported by water stored locally within the streambed sediments (directly beneath the pool) and would be more susceptible to drying through evapotranspiration (less inflow but the same amount of water loss through evapotranspiration).

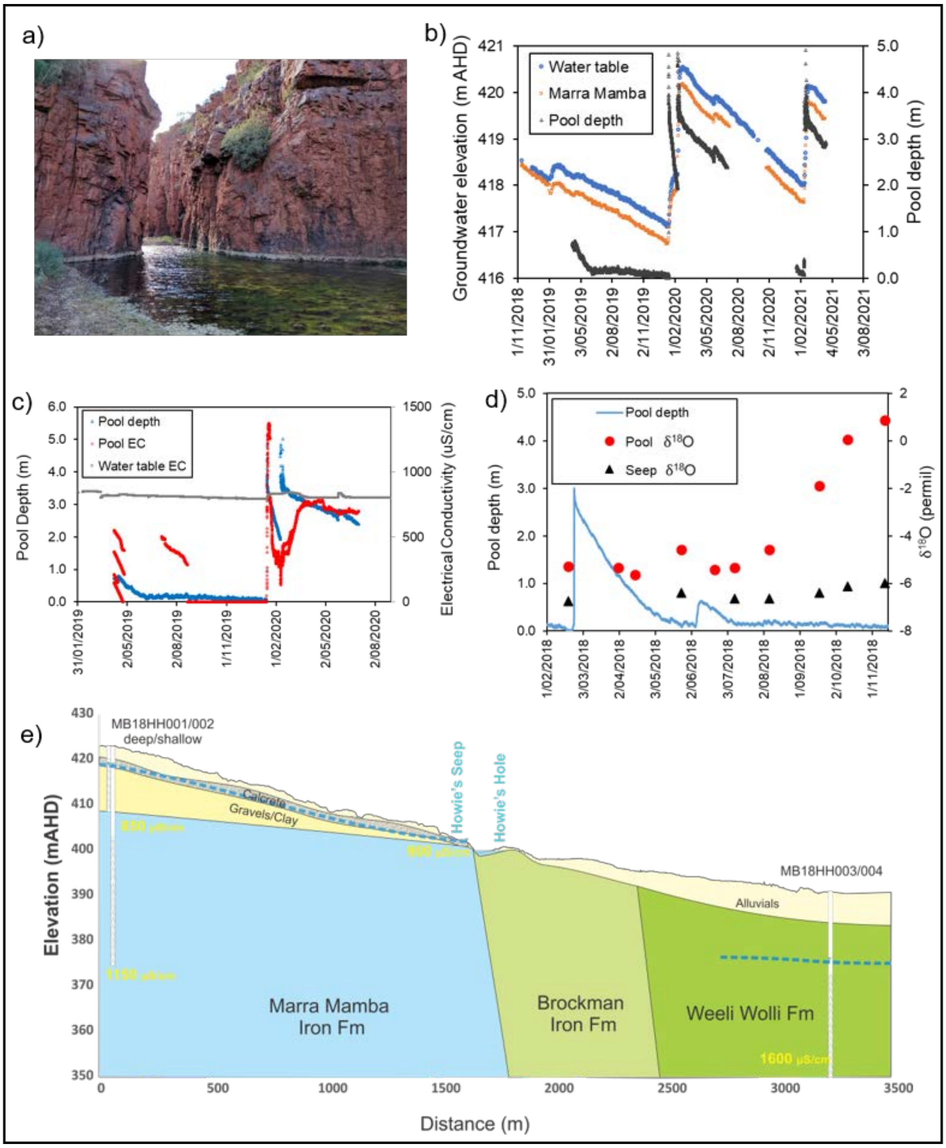

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Figure 9 a) photo of Howie's Hole, b) groundwater elevations and pool depth, c) pool water levels and electrical conductivity of pool and groundwater, d) pool depth and $\delta^{18}O$ showing stable isotopic composition at groundwater seep and evaporative enrichment down-gradient of seep during the dry season, e) conceptual diagram of pool occurrence.

5.2.3 Case study 3: Ben's Oasis

Ben's Oasis is a sequence of three sub-pools that are hydraulically connected during peak water levels and subsequently disconnect during the dry season (Fig. 10a). The pools sit within a major drainage channel that consists of poorly sorted, fine to very coarse (gravel and boulders) unconsolidated alluvial sediments 10's of metres wide and on the order of metres in thickness. The regional water table is within the fractured dolomite of the Wittenoom Formation, which overlies the Marra Mamba Formation. Water levels in the upper pool have been monitored since 2016 and in 2019 a detailed study commenced using environmental tracers to assess the spatial variability of surface water – groundwater interaction along this pool sequence (Chapman, 2019).

Measured pool water levels show consistent seasonal trends with water level spikes of 2-3 m in response to cyclonic rainfall events during summer, followed by approximately five months of relatively steady water levels and then recession over approximately three months (Fig. 10b). These trends are consistent with the water level variation in the adjacent alluvium, which exhibits a similar period of steady water levels then recession following the cessation of summer rains. In contrast, regional groundwater levels increase by about 2 m in response to summer rainfall and then immediately begin to recede. Thus, although snapshot water level measurements indicate that pool water levels are consistent with the regional water table, transient water level data (that includes the water level in the alluvium) demonstrates that inflow of water from within the alluvial sediments within the drainage channel is the dominant driver of water level fluctuations in the upper pool (where the logger was installed). Spatial trends in the persistence of surface water and surface geology are also informative at this site. The regional Wittenoom aquifer is exposed at surface around Pools 2 (some alluvium present) and 3 (no alluvium, just bedrock), but not at Pool 1 (no bedrock, just alluvium). The upper, shallower section of

Pool 1 and Pool 3 dried out as the dry season progressed, but the deeper parts of Pools 1 and 2 persisted throughout the dry-season (during 2019 and 2020). We interpret these spatial patterns of persistence as reflecting evaporation rates (more or less shading by vegetation) and heterogeneity in groundwater inputs (Chapman, 2019).

The results of longitudinal hydrochemical surveys (^{222}Rn and δ^{18}O) along the pool sequence provide an independent line of evidence to validate this interpretation (Fig. 10c). Alluvial water had a ^{222}Rn activity of 17.6 Bq L^{-1} and δ^{18}O of -6.3 ‰. The regional Wittenoom aquifer had a lower ^{222}Rn activity of 8.1 Bq L-1 and more depleted δ^{18}O of –7.26 ‰. At the top of Pool 1, ^{222}Rn activity was 7 Bq L^{-1}. Given that degassing of radon to the surface is rapid and the water level at the time of sampling was shallow, the source of water inflows must have a much higher ^{222}Rn activity than 7 Bq L^{-1} and it is therefore most likely that inflows here are dominated by the higher-Rn alluvial water. Isotopic δ^{18}O values of around –6 ‰, are also consistent with inflow of alluvial water. ^{222}Rn activities then decrease along the pool to around 0.5 Bq L^{-1} (indicating degassing, and the absence of further groundwater inputs) as stable isotopic values enrich to just under –5 ‰ (reflecting evaporation and the absence of further groundwater inputs). Water at the top of Pool 2 had ^{222}Rn of 2 Bq L^{-1} (greater than at the bottom of pool 1) and δ^{18}O of -6.3 ‰ (more depleted than at the bottom of Pool 1). These data indicate further water inflows from the subsurface, along this pool, with a lesser proportion of alluvial water, and more regional groundwater, as well as through-flow from Pool 1 (inferred from relative water levels in the pools). In Pool 3, ^{222}Rn remains around 2 Bq L^{-1} indicating further groundwater inputs, but the stable isotopic values are more enriched (possibly due to the shallow water depth allowing for enhanced evaporation).

Streambed temperatures within the pools were also mapped (temperatures measured every 0.2 - 1 m along transects 1-10 m apart) in early September, when regional groundwater was 29 °C, and alluvial water was 20 °C (Fig. 10d). Measured temperatures were recorded at dawn to reduce the effect of direct solar radiation and pool depth variability (max pool depth was 0.5 m). Streambed temperatures in the pools ranged from 17-23 °C, with the warmest water (>20 °C) at the top of Pool 1, and temperatures between 19-20 °C in middle of Pool 2 and at the top of Pool 3. These results are broadly consistent with the other results, but the approach is likely to be more conclusive in the presence of larger temperature gradients. The application of vertical temperature profiles to infer water fluxes at this site was also limited by the substantial lateral component of the subsurface flow-field (i.e. violating the assumption of 1D flow) and flood events that removed or damaged monitoring infrastructure.

There are two hydraulic mechanisms supporting the water balance and persistence of these pools; alluvial through-flow and regional groundwater discharge (Fig. 10e). Based on these data we conclude that the persistence of Ben's Oasis throughout the dry season is supported by regional groundwater inflows from the unconfined aquifer where it is exposed at surface (see Section 2.3), but the water balance of Pool 1 is dominated by exchange with the alluvial water (see Section 2.2). This importance of the alluvial water storage in supporting the largest of these pools is only evident based on time-series water level data from the alluvium. Given only snapshot water level measurements from the regional aquifer and one location in the pools, the similarity in water level elevations would lead to the conclusion that regional groundwater discharge was the dominant supporting mechanism. The substantial spatial variability captured in the longitudinal hydrochemical survey also highlights the risks of making conclusions about surface water – groundwater interactions from snapshot hydrochemistry measurements in just one location within a given pool or pool sequence.

The persistence of these pools through the dry season is dependent on influx of water from the regional unconfined aquifer. They will therefore be susceptible to groundwater withdrawals from the regional aquifer if they reduce the hydraulic head to below the level of the ground surface at the pools. The water balance of these pools is also controlled by the interaction with water stored in the alluvium (alluvial through-flow). Therefore, the pools are also susceptible to reductions in rainfall or increases in temperature (and evapotranspiration) that reduce the volume of water storage (and therefore water levels) within the streambed alluvium. A reduction in the area of the surface catchment resulting from human activity could also similarly alter the water balance of these pools.

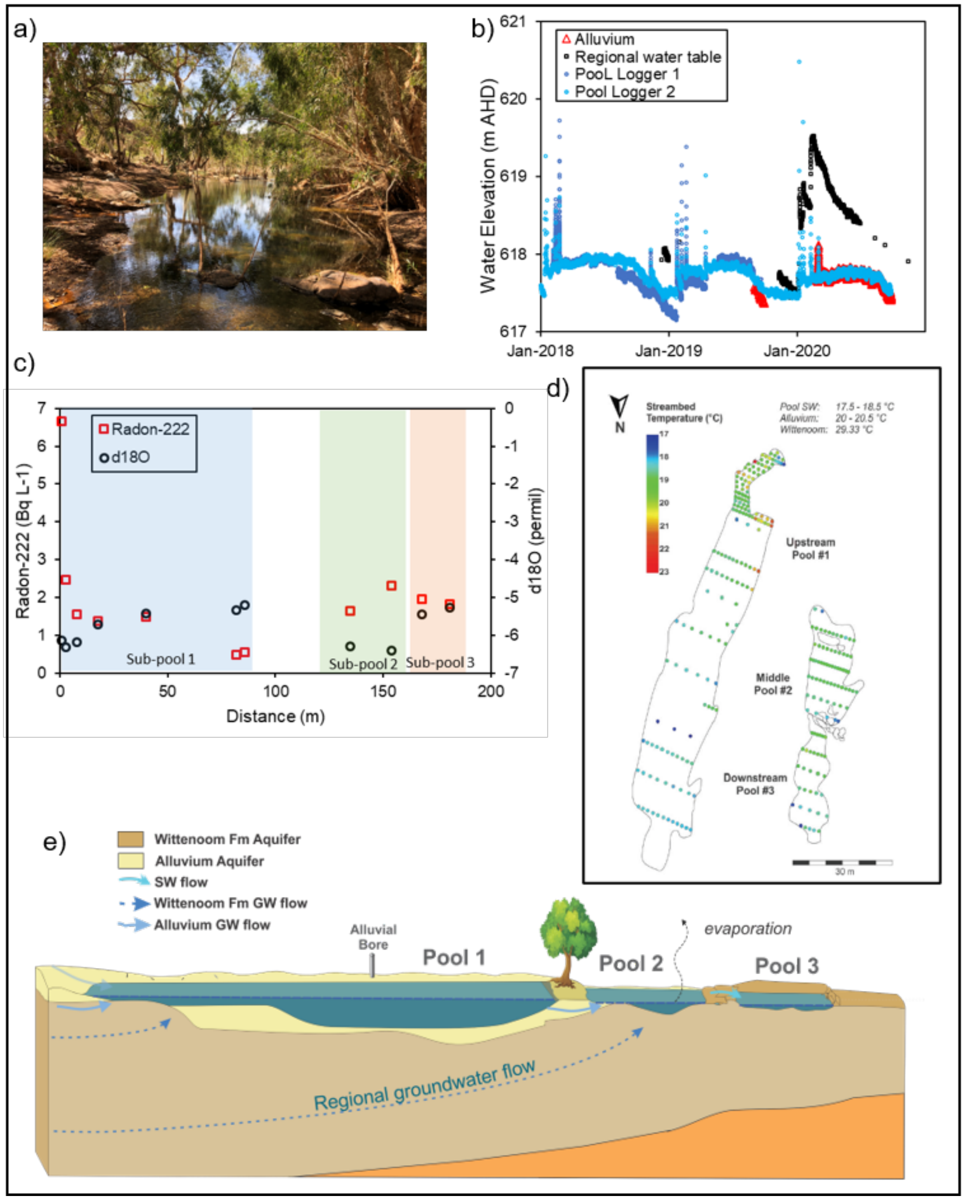

Figure 10 a) photo of Ben's Oasis, b) water levels in the Pool 1 (logger 1 elevation was surveyed, logger 2 elevation was approximated by matching data from logger 1), alluvium (DP1) and regional unconfined aquifer, c) spatial variation in radon activities and $\delta^{18}O$ along the pool sequence, d) temperature mapping of pool sediments and e) conceptual diagram of mechanisms supporting pool persistence.

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6 Conclusion and recommendations

It has now been 100 years since groundwater springs were documented in published literature (Bryan, 1919; Meinzer, 1927; Meinzer, 1923) and while frameworks for groundwater springs and aspects of non-perennial streams (e.g. Costigan et al., 2016) exist, there hasn't yet been a hydraulic classification system defined that applies to persistent in-stream pools. Persistent pools are an important feature along non-perennial rivers and these types of systems are under increasing pressure from altered hydrology associated with shifting climates and anthropogenic activities (Steward et al., 2012). This paper identifies the dominant hydraulic mechanisms that support pool persistence. Each mechanism has varying degrees of connection to groundwater or differing controls on groundwater outflow (geological barrier vs topography). Pools can be supported by multiple hydraulic mechanisms; through-flow pools with some regional groundwater input are likely common, but it can be difficult to definitively identify this regional component of the water balance. Susceptibility to hydrological change depends on the mechanism(s) of pool persistence and the spatial distribution of stressors relative to the pool. While the existing literature hints at the hydrologic and geologic constraints imperative to pool persistence, the framework presented here provides a more scientific characterisation as required to sufficiently understand and protect persistent pools globally. We also present a suite of tools that can be used to test our conceptualism of pool hydrology at a given site, allowing this framework to be applied to the real world.

With limited resources and access to sites, trade-offs must be made between detailed characterization of one pool vs a minimal data set at many pools. Snapshot data from multiple pools at one point in time can help distinguish perched pools vs groundwater discharge pools (i.e. pool water hydrochemically similar or different to rainfall or groundwater), but in some cases water types are difficult to distinguish based on easily measured parameters like electrical conductivity or stable isotopes of water (Bourke et al., 2015). Highly instrumented sites with robust geological mapping, monitoring wells and temporal hydrologic data are required to be confident of pool mode of occurrence. Given limited resources, we suggest that time series water level measurements (groundwater and surface water) and hydrochemistry data from fewer pools is more likely to provide useful insights than snapshot data from many pools. This will be particularly effective if detailed empirical data sets at archetypal sites can be used to group pools based on landscape position or geology.

The study of persistent river pools is a developing science and much remains to be done. Policy makers increasingly require accurate information on the mode of occurrence of surface water pools to put forward management plans to mitigate and/or minimise the adverse impacts of human activities (Leibowitz et al., 2008). This framework is subject to refinement as sufficient data becomes available to fully characterise pool water balances and mode of occurrence. Extension of this framework to facilitate the incorporation of biological and sedimentological processes is also desirable. Persistent river pools exist in all climates across the globe, and consistent data on geomorphology, hydrology and ecology should be collected at multiple features so that generalized patterns and processes can be elucidated. The nutrient and carbon transport between pools during flows and the effects of anthropogenic disruption to groundwater inputs or surface water flushes into these pools is also not well known. These disruptions can be detrimental to water quality if the anthropogenic inputs are contaminated (Jackson and Pringle, 2010), but may also support seasonal connectivity that benefits the ecosystem by distributing nutrients and organic matter between pools (Jaeger et al., 2014). Effects of climate change (e.g. lower groundwater levels, thermal loading, and altered storm cycles) also combine with geomorphological and biological factors to impact ecosystem function, but these mechanisms are not yet well understood.

Data Availability

The data used in Section 5 of this paper are the property of Rio Tinto. Access to these data may be requested by contacting Shawan Dogramaci (shawan.dogramaci@riotinto.com)

Author Contribution

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SB and MS prepared the text of the manuscript with input from all co-authors. PH, SC, SD and SB collected and analysed the data presented in Section 5. PH and SB prepared the figures.

Competing Interests

The authors declare that they have no conflict of interest.

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