Surface Water Groundwater Interaction in Semi-arid Catchments and Implications for Groundwater Management and Reach Classification

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Surface Water Groundwater Interaction in Semi-arid Catchments and Implications for Groundwater Management and Reach Classification

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The University of New South Wales

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In semi-arid environments, groundwater provides the basis for ecosystems and human activity. While groundwater responds to large flow events and recharge is thought to occur predominantly via stream channels (i.e., focused recharge), little is known about its spatio-temporal characteristics and how it relates to effective water resource management.

This study aims to identify the dominant groundwater recharge mechanisms and the extent of resource recovery following a climatic transition from dry to wet in the Maules Creek Catchment, New South Wales, Australia. The dynamic interactions between streamflow and groundwater levels were analysed using 15- to 30-minute resolution data at three representative sites along intermittent and perennial stream channels. At the catchment scale, long-term trends in groundwater level have been analysed using monthly records collected from 35 monitoring bores since the 1980s.

The site and catchment scale analyses demonstrate the following:

(1) In the intermittent section, groundwater level drawdown of up to 5 m was recovered to near or above pre-irrigation (~1980) levels.

(2) Along the intermittent stream reaches, groundwater recharge depends on antecedent groundwater level and soil moisture conditions.

(3) Episodic high stream stage events provide limited recharge to perennial stream reaches since rises in stream stage mainly cause temporal bank storage. Groundwater level increases near these perennial reaches are consequently due to this bank storage and due to loading effects occurring on a time scale of weeks to months.

(4) Areas away from stream channels exhibited an overall decline and the groundwater resource was not restored to pre-abstraction conditions.

This thesis demonstrates that ephemeral and intermittent streams may exhibit significant focused recharge relative to the diffuse recharge over the catchment; whereas, aquifers along perennial streams can only temporarily store water because these streams generally act as groundwater drains. Streams can dynamically behave as ephemeral, intermittent or perennial depending on the longer-term climate-groundwater interaction. Consequently, hydrologic classification of stream reaches in semi-arid areas must allow for dynamic changes and account for the connection to groundwater. This thesis further highlights the potential for managed aquifer recharge along ephemeral and intermittent streams to better utilise stormflow and, therefore, to drought-proof rural communities.

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Abstract

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Content

Figures

Tables

1 Introduction

1.1 Motivation

Improving our understanding and quantification of surface water-groundwater interactions provides a necessary basis for sustainable groundwater management into the future (Winter et al., 1998, Sophocleous, 2002). In semi-arid and arid regions, effective management of water resources is the key to sustainable development, especially during droughts. It has been demonstrated that, unlike in humid environments where diffuse (areal) recharge dominates the groundwater input, in semi-arid and arid regions recharge through stream channels (or focused recharge) is the major source of groundwater replenishment (Lerner et al., 1990, Scanlon et al., 2006, Iverach et al., 2017). However, there is no well-established method to quantify stream recharge, especially during and after floods, and especially where spatio-temporal monitoring is scarce. It is common in water balance modelling to assume fixed recharge rates in time and space, and this results in considerable uncertainty in proportioning recharge to all components of the water budgets and estimating the sustainable yield of an aquifer (Giambastiani et al., 2012, Iverach et al., 2017). For example, if too much recharge is attributed to diffuse recharge, then the contributions to groundwater recharge from flood waters or river leakage will be underestimated.

Groundwater recharge is traditionally defined in the literature as the quantity of the downward flow of water that reaches the groundwater table (Freeze and Cherry, 1979, Lerner et al., 1990, Healy and Scanlon, 2010). However, for recharge near stream channels, groundwater cannot necessarily be regarded as an isolated reservoir, but, rather, a dynamic component interacting with the stream on different timescales. Therefore, recharge should be studied in the context of a relevant timescale. Flood water may enter and leave the subsurface as bank storage on a timescale of days to months, while recharge to regional groundwater flow may take years to millennia to later discharge into streams (Tóth, 1963). To advance our knowledge of recharge processes the dynamics involved in hydrological connectivity requires significant development in both field measurements and numerical simulations (Bracken and Croke, 2007, Guzmán et al., 2016, Ward et al., 2017).

Considerable research on stream-aquifer interaction during high stream stage events has been undertaken in perennial systems where water saturated hydraulic connection exists between the stream and the groundwater (Barlow et al., 2000, Chen and Chen, 2003, Ha et al., 2008, Maharjan and Donovan, 2016). Few studies have reported on disconnected systems where the hydraulic link becomes variably saturated or even dry because the groundwater level falls below the base of the stream channel (Vázquez-Suñé et al., 2007, Shanafield et al., 2012). Unlike connected systems where the aquifer response to the stream stage is largely governed by the hydraulic diffusivity, a disconnected aquifer response is affected by channel width and depth, streambed sediment thickness as well as the lateral and vertical hydraulic conductivity distribution (Brunner et al., 2009). In addition, the antecedent condition of the system, such as depth to the groundwater table, soil moisture content, and the initial degree of disconnection, affects the magnitude of recharge that enters an aquifer through the streambed. However, in many groundwater models, antecedent conditions are often assumed constant or simplified (Crosbie et al., 2008). In particular, the hydrological classifications of streams, as either ephemeral, intermittent or perennial, are a static classification defined and based mainly on flow statistics.

The Millennium Drought, which occurred in most of southern and eastern Australia between 2001 and 2009, has been considered one of the worst since European settlement (Leblanc et al., 2009, van Dijk et al., 2013). It has been attributed to above-average air temperature with record-low rainfall (Bond et al., 2008). At the end of the drought, a change in the climatic regime brought above-average rainfall and floods between 2010 and 2013. This wet run initiated the recovery of groundwater levels in many catchments around Australia (King et al., 2014). It is therefore a common notion among the general public that such floods provide sufficient recharge to replenish groundwater resources, which might then be utilised in future dry years. Newspaper articles with titles such as "Aquifers lap up the water" (Brown, 2011) "A land recharged" (Browne, 2012) and "Water reserve topped up" (Williams, 2013) were common from mainstream media outlets in Australia. However, a few wet years following prolonged drought may not reverse the overall downward trend, especially when the depletion is related to both natural drought and groundwater pumping for agricultural and domestic consumption (Chen et al., 2016). Although small or shallow aquifers are likely to be fully recovered from the preceding drought, relatively larger or deeper aquifers require a longer period of rain and flooding to be fully recharged (NWC, 2011). To better inform water resource management and to educate the public, it is crucial to identify and to quantify the processes controlling recharge and aquifer recovery. This thesis uses high resolution streamflow and groundwater hydrograph data to advance our knowledge of when and how the aquifers are recharged along an inland river corridor in a semi-arid environment, as well as the changes under climate variation. Such insights are useful for informing efficient water resource allocations under an integrated water resource management framework and may assist irrigation farming to move toward a sustainable future. Moreover, it might enable better management of groundwater-dependent ecosystems in dryland environments and, consequently, promote a healthy aquatic environment and good water quality.

1.2 Study objectives

When a net hydraulic gradient is maintained towards the aquifer, it is observed that flooding replenishes groundwater resources near river channels. The main objective of this study is to explore the relationship between the extent of stream recharge and the available groundwater storage capacity under the stream channel. It is hypothesised that, rather than separate the conceptual model of connected and disconnected streams, the connection state is dynamic and can transform from one to the other over time and in space. This will be demonstrated through investigating the hydrograph response to flow at representative sites with different degrees of connection.

More specifically, this study aims to

- analyse the groundwater responses to high stream stage maintained for days to weeks;
- describe specific conditions that may promote recharge;
- highlight the implications of these condition for water resource management over the long term; and
- assess the appropriateness of the hydrological stream reach classification in the context of water management.

This study is based on groundwater and surface water level monitoring data collected at Maules Creek Catchment in the Namoi Valley. High frequency (15-min. interval) data have been collected from three sites along a hydrologic and geomorphic gradient from the mountain front to the floodplains. This data is used for event analysis at the field site scale (~ 100 m) to identify favourable recharge conditions. Long-term monitoring data collected by the NSW Office of Water with wide spatial coverage over the entire alluvial part of the catchment is used for statistical analysis of past natural climatic transitions from dry to wet in order to reveal the controlling recharge mechanisms.

1.3 Background and literature review

1.3.1 Developments in the field of surface-water/groundwater interactions

As a part of the hydrologic cycle, groundwater enters and leaves the subsurface in the form of recharge and discharge (Freeze and Cherry, 1979). The connection between discharge, surface flow and precipitation was observed and acknowledged by ancient Greek philosophers (Deming and Fetter, 2004). Nevertheless, the hydrology of groundwater and surface water differ significantly on both spatial and temporal scales and, thus, has traditionally been studied and managed separately (Winter et al., 1998, Braaten and Gates, 2003).

The concept of surface water-groundwater interactions acknowledges different groundwater flow regimes on local, intermediate, and regional scales due to topographical, geological, and climatic characteristics of a catchment (Tóth, 1970). Research on surface water-groundwater interactions has evolved over recent decades to become more holistic through the last few decades by consideration of the surface and subsurface to become one connected system (Sophocleous, 2002). Studies have been expanded from riverine systems to mountain, coastal and karst terranes (Winter, 1995). More recently, the ecological complexity, biodiversity and local endemism that are contained in groundwater ecosystems have also been recognised (Humphreys, 2009). The recharge of aquifers not only provides long-term storage of stable water supplies, but also sustains groundwater-dependent ecosystems (Datry et al., 2005).

1.3.2 Stream classification

Conceptualisation of natural surface water-groundwater interactions has been evolving for decades (Brunner et al., 2011). Meinzer (1923) described the stream as either *gaining* or *losing* water to its underlying groundwater aquifer. A gaining stream receives groundwater flow from adjacent aquifers (i.e., discharge, which is also known as baseflow in hydrology). For this to happen, the surface water level of the stream needs to be lower than the groundwater table. Conversely, a losing system is one in which water flows out of the stream and into the aquifer where the groundwater table is lower than the surface water level of the stream (i.e., focused recharge). These systems are referred to as *connected* since the saturated zone (i.e., groundwater) is hydraulically connected with the river channel (Walton, 1955). If the groundwater table is sufficiently deep, then any changes in the groundwater level will not alter the infiltration rate. This scenario is referred to as a *disconnected* system (Moore and Jenkins, 1966, Peterson et al., 1984). In contrast to the disconnected stream is the *insulated* system described in Meinzer (1923), which is separated by a clogging layer or aquitard and, therefore, contributes nor receives water from the zone of saturation.

Winter et al. (1998) summarise the terminology and hydrodynamics of surface water-groundwater interactions (illustrations shown in Figure 1.1). Although these stream classification categories provide a basic understanding, they are often too simple for describing natural streams, especially in semi-arid and arid environments. Winter et al. (1998) also noted that a stream may have a combination of gaining, losing or disconnected reaches, as the flux direction may vary over different geomorphological units, or timescales in response to seasonal and climate variations. Therefore, along a stream, the flux direction may vary both spatially and temporally. Yet, in catchment scale modelling practice, it is still a numerical challenge to evaluate whether a system is connected or disconnected. In particular, for dryland rivers with unsaturated zones, the linearity of hydraulic coupling terms involved in the unsaturated flow processes within disconnected systems involves significant uncertainty (Costa et al., 2012). For complex streams, ambiguous classifications such as variably connected–disconnected and variably gaining/losing are sometimes used (Ivkovic, 2009).

Waterways are often classified as perennial, intermittent or ephemeral. For large rivers with multiple tributaries and large catchment areas, flow is generally maintained throughout the year, except in extremely dry conditions. Such streams are referred to as *perennial*. At the other extreme, e*phemeral* streams only flow during or directly after rainfall events or in response to snowmelt (Stringer and Perkins, 2001). Since ephemeral streams are disconnected from the groundwater table, they receive no baseflow contribution. Intermittent streams generally flow in the wet season and cease flowing in the dry season (Uys and O'keeffe, 1997, Anna et al., 2009).

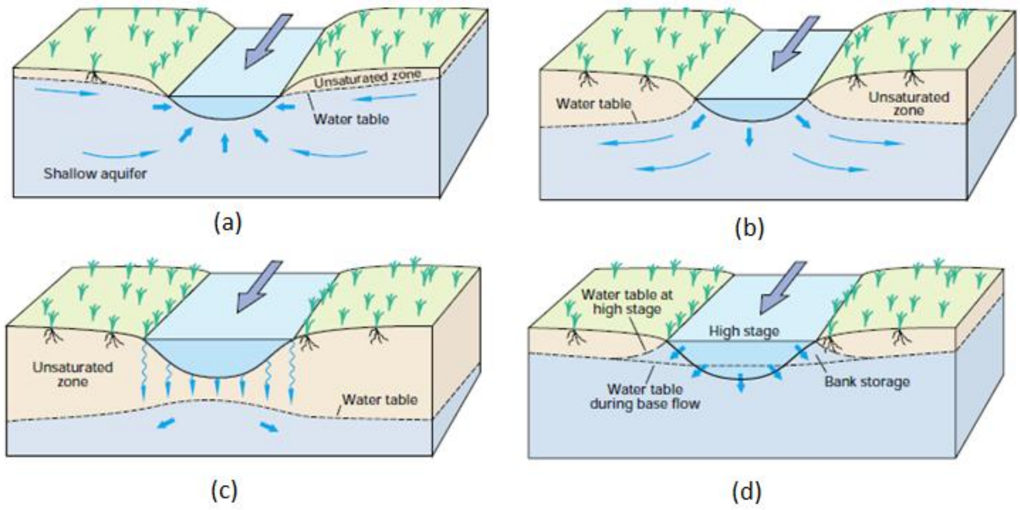


Figure 1.1 Scenarios of surface water-groundwater interactions: (a) gaining stream, (b) losing stream, (c) disconnected stream, and (d) bank storage. Adapted from Winter et al. (1998).

1.3.3 Disconnected reaches and water infiltration

In a disconnected reach, the stream is hydraulically separated from the aquifer by an unsaturated zone. Water first infiltrates vertically through the unsaturated zone and, therefore, raises the groundwater table directly below the stream. This local zone of higher water level is referred to as a *groundwater mound*. The mound is maintained by vertical infiltration and decays when the stream ceases to flow. The groundwater moves laterally away from the stream to form focused recharge. In the case of disconnected stream (i.e., where the water table is deep), the infiltration rate entering the streambed is at its maximum and the saturated zone increases vertically and laterally (Osman and Bruen, 2002). With a decreasing water content in the underlying unsaturated aquifer, the head difference between stream and groundwater no longer affects the downward infiltration rate (Brownbill et al., 2011). Instead, the infiltration rate is controlled by the variable saturated hydraulic conductivity of the sediments. Brunner et al. (2009) used numerical modelling to show that, for a given aquifer thickness and stream width, the depth required for disconnected conditions depends on channel depth, streambed sediment thickness, as well as the lateral and vertical hydraulic conductivity. To further define how changes in the groundwater table affect surface water resources, Brunner et al. (2009) defined a transition flow region where the semi-conductive streambed sediment intersects the capillary zone but not the groundwater table (Figure 1.2).

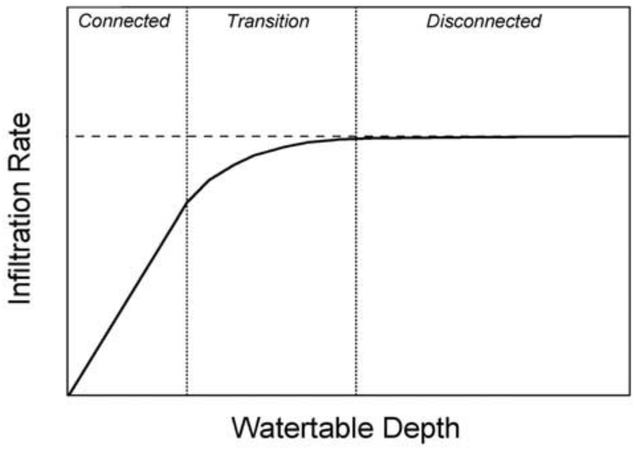


Figure 1.2 Stream water infiltration rate as a function of groundwater table depth (Brunner et al., 2009).

1.3.4 Connected reaches and bank storage

In the connected stream-aquifer system, the flux between stream and aquifer is controlled by the head difference and the aquifer hydraulic conductivity, as described in the original work conducted by Henry Darcy, as reflected in the following equation:

$$Q = -KA\frac{dh}{dL}$$

This is commonly known as Darcy's Law (Freeze and Cherry, 1979). Transmissivity (T) measures under the hydraulic gradient of 1, the volume of water that can move horizontally through a unit width of the saturated zone.

$$T = Kb$$

In an unconfined system, storativity is approximately equal to specific yield:

$$S = S_y$$

If the water level in the stream fluctuates, the aquifer response is controlled by the hydraulic diffusivity of the system. This is defined as

$$D = \frac{T}{S}$$

For a homogeneous, isotropic aquifer, hydraulic diffusivity can be determined using the delay time response in the aquifer to fluctuations in stream stage (Pinder et al., 1969). For non-idealised field studies, aquifer response time can provide information about the aquifer properties (Sophocleous, 2011, Kelly et al., 2013, Graham et al., 2015, Rau et al., 2017).

Bank storage is the amount of flood water that is temporarily held in bank sediments (Hantush et al., 2002). During a rapid rise in stream stage caused by events such as storm precipitation, rapid snowmelt, or release of water from an upstream reservoir, water flows from a stream into the stream banks and underlying sediments due to the head difference between the river and the groundwater in the surrounding area (Winter et al., 1998). As the flow is impeded by the porous medium, only a small portion may infiltrate further sidewards and raise the groundwater level in the surrounding aquifer. Once the stream level falls, most of such gain in the proximity of the stream channel discharges back into the river as return flow (Figure 1.1 [d]).

It is difficult to distinguish bank storage from groundwater recharge. At the scale of individual events, analytical solutions based on hydrograph analysis have been proposed for separating bank storage from continuous baseflow (Cooper and Rorabaugh, 1963, Hunt, 1990, Barlow and Moench, 1998). Simpson et al. (2013) observed through chemical and isotopic analysis that flood water dominated baseflow for extended periods following large floods. This suggests that the flood recharge is released to steam after the flood. When bank storage discharges to a stream as return flow, tracer studies can qualitatively identify the bank storage from baseflow (Cartwright et al., 2014). As the bank storage water resides in the aquifer for some time, the difference between bank storage and recharge is somewhat arbitrary and the distinction depends on the time frame selected for the analysis.

1.3.5 Aquifer responses

Through numerical modelling and field data analysis, Shanafield et al. (2012) concluded that the disconnection status between the stream and aquifer can be determined when a change in the aquifer head is greater than the change in the stream stage. This theory was tested using field data observed for disconnected reaches. However, Shanafield et al. (2012) also acknowledged that the magnitude of the disconnected groundwater response does not scale linearly with change in aquifer parameters, and that further field-based studies are required to develop convenient and reliable methods to determine connection status using the transient aquifer hydrograph response.

Following streamflow, the recession characteristics of the groundwater level enable groundwater table fluctuation analysis to quantify focused, indirect recharge over both the long-term and event time scales. Cuthbert et al. (2016) showed that groundwater head responses to ephemeral streamflow events are controlled by pressure redistribution after water has entered the subsurface. This redistribution process operates on the timescale of days to weeks in the vertical direction (below the stream), weeks to months in the transverse direction (perpendicular to the stream), and years to decades in the longitudinal (parallel to the stream) direction (Rau et al., 2017). A rise in the groundwater table can also be the result of the hydrostatic loading effect, which is indicated by a swift rise in the groundwater level along with the increase of the stream stage, then followed by a relatively fast decline in head as the flood recedes (Doble et al., 2012).

1.3.6 Groundwater recharge

Groundwater recharge is defined as the process whereby surface water or rainfall flows downwards from a surface water body or the unsaturated zone, providing an inflow to the saturated zone of an underlying aquifer. This inflow raises the groundwater table and causes an increased storage of water (Freeze and Cherry, 1979, Lerner et al., 1990, Healy and Scanlon, 2010).

Diffuse (or areal) recharge is distributed over the land mass (Healy and Scanlon, 2010). When precipitation occurs, diffusely infiltrating water increases the soil moisture content. For groundwater recharge to occur—with the exception of bypass flow through preferential flow paths formed by rootlets or cracks—the moisture content of the soil profile needs to exceed the field capacity of the sediment and be maintained above field capacity for a period long enough for water to migrate below the root zone. Otherwise, the water may be transported back into the atmosphere via evaporation or transpiration. In sub-humid environments, it is estimated that the minimum rainfall required to initiate infiltration should be 2.5 mm/day for more than 3 days (Herczeg et al., 1997). Studies by Markowska et al. (2015) show such threshold is 13 mm in semi-arid environments. Moreover, it was observed in a semi-arid area that increased precipitation can lead to higher water usage through transpiration due to subsequent plant growth (Scanlon et al., 2006).

Focused recharge, on the other hand, is the flow from permanent or temporal surface water bodies to the groundwater table (Scanlon et al., 2006). Mountain-front and mountain-block recharge have been defined as water entering adjacent inter-mountain basin-fill aquifers (Wilson and Guan, 2004). However, when surface water and groundwater interactions are considered, groundwater can no longer be regarded as an isolated reservoir, but, rather, a dynamic component of the hydrologic system (Winter et al., 1998).

Although hydrograph analysis can demonstrate fluctuation in groundwater levels and reveal groundwater recharge, it is difficult to separate the downward mass flux of water from the flood wave, piston flow and loading effects characteristic of deeper aquifers. McDonnell and Beven (2014) emphasised the importance of the combination of hydrograph and tracer analysis, as this can expose the origin of water. Yet, unless tracer data are collected routinely to build a time series with a reasonable temporal resolution, it only provides snapshots of the recharge characteristics. Moreover, when bank storage and hyporheic exchange are involved, distinct chemical or isotopic signatures between surface water and groundwater are often mixed and, therefore, difficult to separate (Wels et al., 1991).

Along river channels, Scanlon et al. (2006) demostraed that preferential and focused flows form important recharge pathways. In an arid environment, Harrington et al. (2002) revealed that flood flows from ephemeral streams lead to a groundwater recharge rate that is ten times higher than the average for the remainder of the catchment. Along a losing river reach in the lower Namoi alluvial aquifer in NSW, observation wells within 10 km of major streams respond noticeably to stream flooding, and the response varies among lithological units in the valley fill (Kelly et al., 2013, Kelly et al., 2014). Iverach et al. (2017) demonstrated with isotope studies that the flood recharge is focused along the river corridor in the lower Namoi. If there is an unsaturated zone underneath a stream, the magnitude and frequency of recharge is dependent on the amount of water that infiltrates the stream bed (Abdulrazzak et al., 1989, Masoud et al., 2013, Shanafield and Cook, 2014).

For streams that are hydraulically connected to nearby aquifers, it is common in arid and semi-arid regions to have rivers that alternate between gaining and losing conditions along their length (Rushton, 2007). King et al. (2014) applied an integrated hydrogeological and hydrochemical approach to analyse the groundwater response to droughts and floods in an alluvial aquifer of the Cressbrook Creek Catchment in Queensland. The study indicated that intermittent streams may change from losing to gaining conditions depending on groundwater levels in the adjacent aquifer. Therefore, a suitable temporal scale is required when quantifying flood recharge for water balance purposes. Rau et al. (2017) used the combination of streambed thermal signatures and water levels to characterise the dynamics of surface water-groundwater interactions in dryland streams. The study suggests that the conditions for any studied reach may switch from a dry stream channel to surface flow sustained for months and back to isolated pools and dry conditions over relatively short periods of time.

1.3.7 Drought and groundwater abstraction

Many studies have investigated the responses of alluvial aquifers to climate variability using water budget methods (Healy and Scanlon, 2010). Thiery et al. (1993) applied a water balance model to 13 years of manually monitored piezometric levels to reveal that a slow aquifer response made it impossible to identify a unique set of simulation parameters.

Where surface water is used on irrigated farms, the groundwater table may rise due to recharge induced by over-irrigation. The rising groundwater table due to water diversions across catchment boundaries has been of great concern in parts of Australia over the past few decades since it may remobilise salts form areas of dryland salinity. Yet, where groundwater is used for irrigation, the amount of drawdown is usually much greater than the magnitude of deep drainage (Hughes et al., 2011, Giambastiani et al., 2012, McCallum et al., 2013).

2 Catchment context and data collection methods

2.1 Overview

This study is based on field sites located in the Maules Creek Catchment in northwestern New South Wales. Located on the western slopes of the Great Dividing Range, the 1,475 km^2 catchment is bound to the north and the east by the New England Fold Belt and the Nandewar Range (Andersen and Acworth, 2009). The Namoi River enters the study area from the south at Gins Leap Gap, which is a narrow constriction (< 2 km) of the alluvium between the low-ridge hills of Pilliga Forest and the Leard State Forest. It is believed to be controlling groundwater flow entering the catchment from the south (Crawford et al., 2009, Giambastiani et al., 2012, McCallum et al., 2013). To the north, the extent of this study ends where the Bibbla Creek enters the Namoi River. The upstream and downstream boundaries were chosen to include the government operated stream gauges at Boggabri (BOG) and Turrawan (TUR; Figure 2.1).

The point of highest elevation is Mt. Kaputar (1506 m on the Australian Height Datum [AHD]) in the Nandewar Range. Away from the ridgelines, the topography is gently undulating with the ground slope generally less than 10%. The topography of the alluvial plains is gently sloped from approximately 340 m AHD in the mountain front area to 230 m AHD near the Namoi River over a distance of approximately 20 km.

2.2 Geology

The geology of the catchment area and the surrounding area has been described in Andersen and Acworth (2009). The underlying Permian and Jurassic claystone, sandstone and conglomerates of the Gunnedah Formation were incised by high energy streamflow in the late-Cretaceous and formed a paleovalley up to 120 m deep (Wallis, 1971, Pratt, 1998). The sedimentary deposition and its texture in the valley were controlled by the paleoclimatic conditions and the amount of water moving through the landscape at the time of deposition (Nichols, 2009). Since the mid-Miocene, coarse sediments have been transported from the headwaters located on the western margins of the Great Dividing Range and deposited in the valley (Acworth et al., 2015). Refer to Figure 2.2 for details of regional geology.

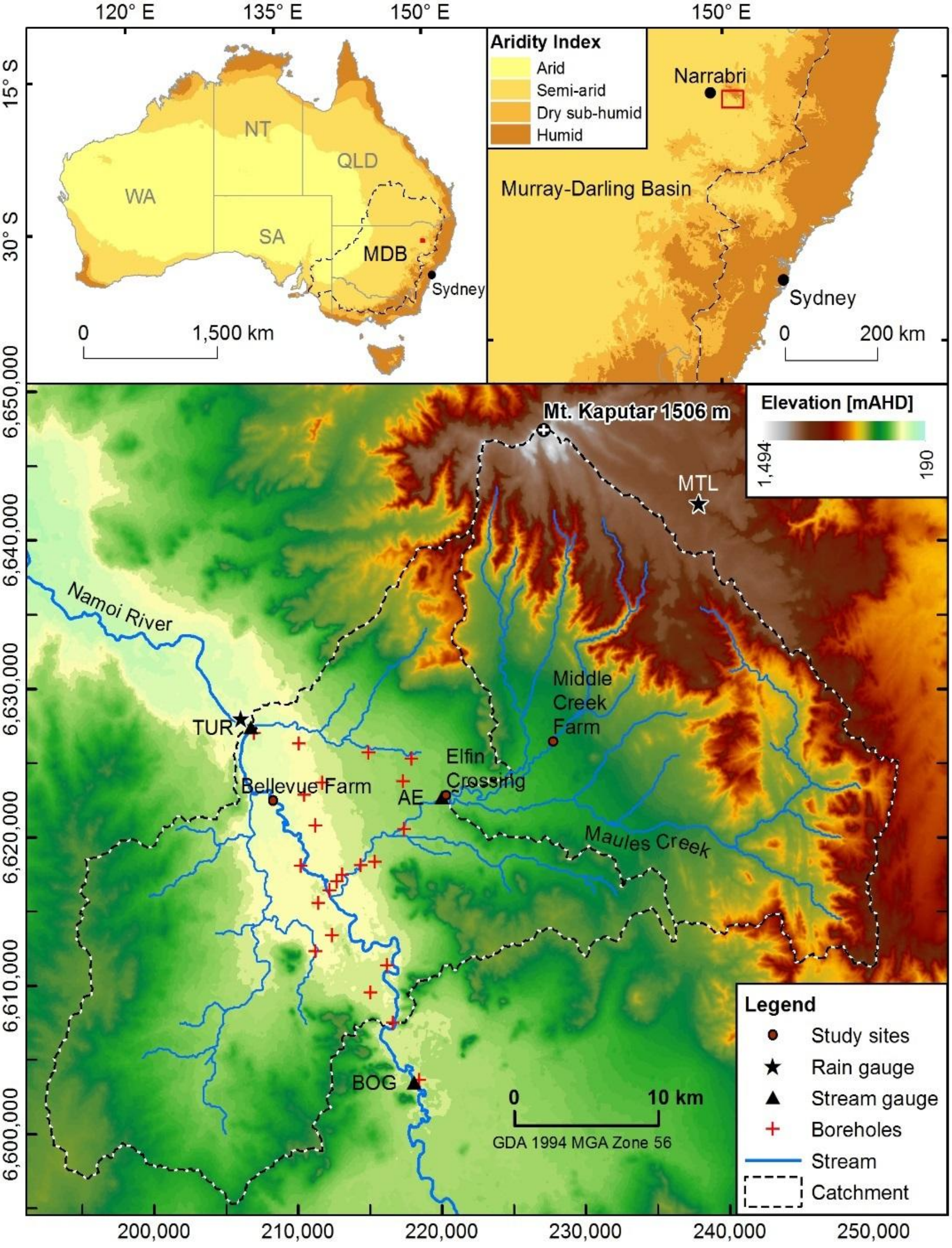


Figure 2.1 (Top) The Maules Creek Catchment located in New South Wales (NSW), Australia is a sub-catchment of the predominantly semi-arid Murray-Darling Basin; (Bottom) locations of three UNSW study sites and government long-term monitoring boreholes. Also shown on the map are the locations of stream gauges

Avoca East (AE), Boggabri (BOG) and Turrawan (TUR) as well as rain gauges at Mount Lindsay (MTL) and Turrawan (TUR).

Kelly et al. (2014) demonstrated with pollen records that the wetter climate with a relatively high-energy depositional environment slowly transformed to a low-energy environment under a dryer climate. The alternating layers of fine-grained materials, especially the present day silty and clayey vertosols on the surface, are believed to be reworked aeolian and alluvial deposits (Young et al., 2002, Wray, 2009). A paleovalley 10 to 15 m deep was also cut along the upstream section of Maules Creek and its tributaries by high energy flows from the Nandewar Ranges. Deposited materials in the current Maules Creek channel are predominately coarse-grain sand and gravel deposits (Rau et al., 2017).

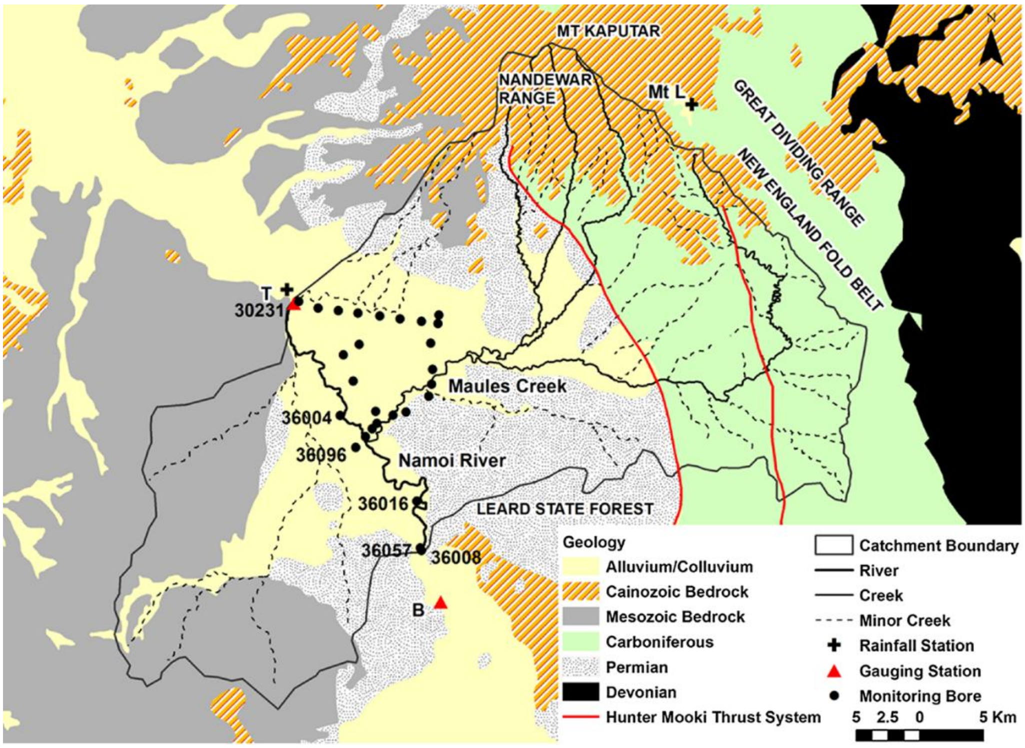


Figure 2.2 Surface geology of the catchment with the locations of river flow gauging stations and the groundwater observation bores (McCallum et al., 2013).

2.3 Hydrology

Major streams in the study area include the Namoi River and Maules Creek. The Namoi River is a 7th order tributary of the Barwon River under the Strahler stream ordering system (Strahler, 1957). Upstream from the study area, the Namoi River drains an area of approximately 22,600 km^2, with an average annual flow volume of 810 GL recorded at Boggabri (BOG) gauge (NSW DI Water, 2015). Three upstream dams regulate between 40% and 90% of the flows in the headwater area (CSIRO, 2007).

Maules Creek is an unregulated 6^{th} order tributary, which drains westward into the Namoi River. Other than the headwaters in Mount Kaputar National Park, it mainly flows through land that has been fully or partially cleared for farming and grazing. Riparian vegetation remains along the channel banks in a zone between 50 m to 300 m wide. Maules Creek is largely intermittent with a perennial section near the Avoca East (AE) gauge (Andersen and Acworth, 2009). There are flows approximately 94% of the time at the AE gauge (Figure 2.1), with an average annual streamflow of 18 GL from the 655 km^2 upstream catchment. Following large storms, high volume surface water flow from the Nandewar Ranges can contribute significantly to the Namoi River (Green et al., 2011). The maximum flow recorded in February 2012 was 41 GL within 24 hours (NSW DI Water, 2015), which was approximately 5% of the Namoi River's total streamflow for that year.

The depth of the groundwater table across the catchment varies between 4 to 15 m below the ground surface, which depends on local hydrological and topographical features. Most productive aquifers are located in the deeper, more coarse-grained part of the main paleovalley beneath the Namoi River. These aquifers contain good quality groundwater and allow for high extraction rates (McCallum et al., 2013). The variable depositional environment of the past formed the current highly heterogeneous sediment, which complicates the understanding of groundwater flow, recharge pathways and groundwater management (Kelly et al., 2013, Kelly et al., 2014).

2.4 Climate

Based on long-term records (1889–2015) acquired through BOM for the TUR weather station (see Figure 2.1 for location), the mean annual precipitation is 606 mm/a while the mean annual potential evapotranspiration is 1526 mm/a. Aridity index, which is the mean annual precipitation divided by the mean annual potential evapotranspiration, is 0.397 for the Maules Creek Catchment. According to the climatic classification by Trabucco and Zomer (2009), the semi-arid climate category falls in between an aridity index of 0.2 and 0.5. From a global perspective, similar semi-arid and arid regions are found in most parts of Australia, central Asia and the Middle East, northern and southern areas of Africa and parts of the western Americas. These regions constitute approximately 40% of the planets land surface (Gamo et al., 2013).

Precipitation patterns feature spatial variation and seasonality. Due to the orographic effect, higher rainfall is observed in the mountain ranges to the east of the catchment. At the MTL station (see Figure 2.1), the mean annual precipitation over the entire period between 1889 and 2015 is 1002 mm/a. However, rainfall in summer accounts for approximately one third of the annual precipitation for both the mountain area and the plain.

2.5 Land development and water use

During times of drought, the natural precipitation is insufficient to grow crops, which are irrigated using either surface water or groundwater. Since the late 1970s, the intensive irrigated farming of cotton, sorghum and wheat has been developed on the fertile alluvial floodplain. As reported by the CSIRO (2007), approximately 95% of groundwater extraction in the Namoi region is used for sustaining intensive agriculture. In years with moderate precipitation and, hence, low streamflow, groundwater pumping from the alluvial aquifer system provides over a third of the irrigation water use (e.g., 36% in the 2000/01 water year)—the remainder is surface water from the Namoi River. During dry years when the river flow is low or is dry, the proportion of groundwater can increase to approximately two-thirds (e.g., 65% in the 2003/04 water year) (CSIRO, 2007).

Water use in the study area is currently regulated under the *Water Sharing Plan for the Namoi Unregulated and Alluvial Water Sources* and *Upper and Lower Namoi Groundwater Sources* (NSW DPI, 2018). As an outcome of the national water reforms in the 2000s, the water sharing plan aims to deal with the over-allocation and over-extraction of the groundwater resource (Kuehne and Bjornlund, 2006). Under a water sharing plan, each catchment is allocated a quantity of water available for use over a specific period. The available water is allocated between the consumptive—via water access entitlements—and the non-consumptive pool, such as for environmental use (Hayball, 2010). For each groundwater source, the extraction limit is based on recharge estimated under the assumption of static diffuse recharge by rainfall infiltration, leakage from the rivers, irrigation deep drainage and inter aquifer flow (NSW DPI, 2016).

2.6 Monitoring programs and data collection methods

Groundwater levels have been monitored by the NSW Office of Water in irrigation areas. The monitoring program managed by the NSW Office of Water is primarily aimed at recording changes in areas of large-scale groundwater extraction in major irrigation areas. Water levels in 35 piezometers in the Maules Creek Catchment were manually dipped since the mid-1970s (NSW DI Water, 2015). The measurement interval was approximately monthly at the beginning of the data series, but decreased to between every 4 to 8 weeks since the mid-1980s (Kelly et al., 2013). Since 2005, pressure transducers have been installed in selected boreholes that are programmed to sample at an hourly frequency.

Groundwater levels at selected locations across the catchment have also been recorded by UNSW for research purposes. The monitoring infrastructure has been established in stages since 2007, supported by funding from the Cotton Catchment Communities CRC, the Cotton CRDC and Commonwealth NCRIS funding with the specific aim to further the understanding of surface water-groundwater interaction. The data has been collected via automated pressure transducers at a high frequency of 15 minutes. Pressure measurements were corrected for barometric influences either by using vented loggers or manually using barometric pressure time series recorded with specific transducers or weather stations.

Surface water levels are monitored by the NSW Office of Water at Avoca East (AE), Boggabri (BOG) and Turrawan (TUR) gauges. Stream levels collected at the AE gauge are used without further processing. In addition, stream levels (uncalibrated gauges) were measured using pressure transducers at Middle Creek Farm (MC) and Bellevue Farm (BV) for shorter time periods between 2014 and 2017. These pressure transducers were placed in perforated PVC pipes and anchored in the streambed using star pickets (see **Figure 2.3** for example). Unfortunately, a number of such monitoring points were lost during high-energy storm flows. The remaining records from MC were installed on a sand/gravel bar to the east of the stream centreline. As a result, it only measures changes in stream level above an elevation of 300.9 m.

Figure 2.3 Monitoring stream level at a site similar to Middle Creek Farm.

3 Transect scale analysis of key monitoring sites

Groundwater level responses at three locations along a hydrologic and geomorphic gradient from the mountain front to the floodplains were recorded and analysed. This includes a disconnected, intermittent stream section at Middle Creek Farm (MC) in the upper catchment; a connected, perennial stream section at Elfin Crossing (EC) in the mid-catchment; and a perennial to intermittent section of the Namoi River at Bellevue Farm (BV) in the developed river floodplain (lowest point of the catchment); see Figure 2.1.

High stream level events in Maules Creek and its tributaries were typically triggered by storm runoff from the Nandewar Ranges. Prior to the March 2014 storm, the catchment experienced a dry summer (Figure 3.1). For 120 days, a threshold of 20 mm in daily rainfall was exceeded only in the mountain ranges in February 2014. Streamflow events in response to rainfall were entirely absent from all three sites during this time. Between the 24th and 28th of March 2014, over 180 mm daily rainfall was recorded by the stations on the floodplain (TUR) and in the mountains (MTL). The rainfall record at the Mount Kaputar weather station illustrated that a typical stream runoff required to drive flow into Maules Creek requires a cumulative rainfall in excess of 150 mm over a 7-day period. Such a rainfall event was higher than the 50% Annual Exceedance Probability (AEP) at the monitoring location (BOM, 2015, BOM, 2016).

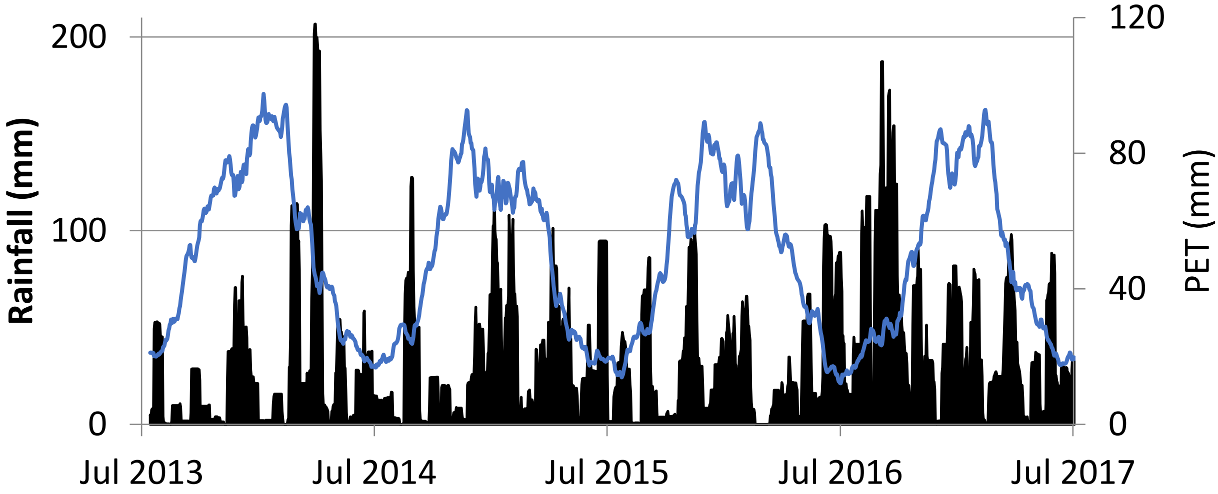


Figure 3.1 14-day running sum of rainfall (bar) and PET (blue line) at Mt Lindsay (MTL) between 2013 and 2017.

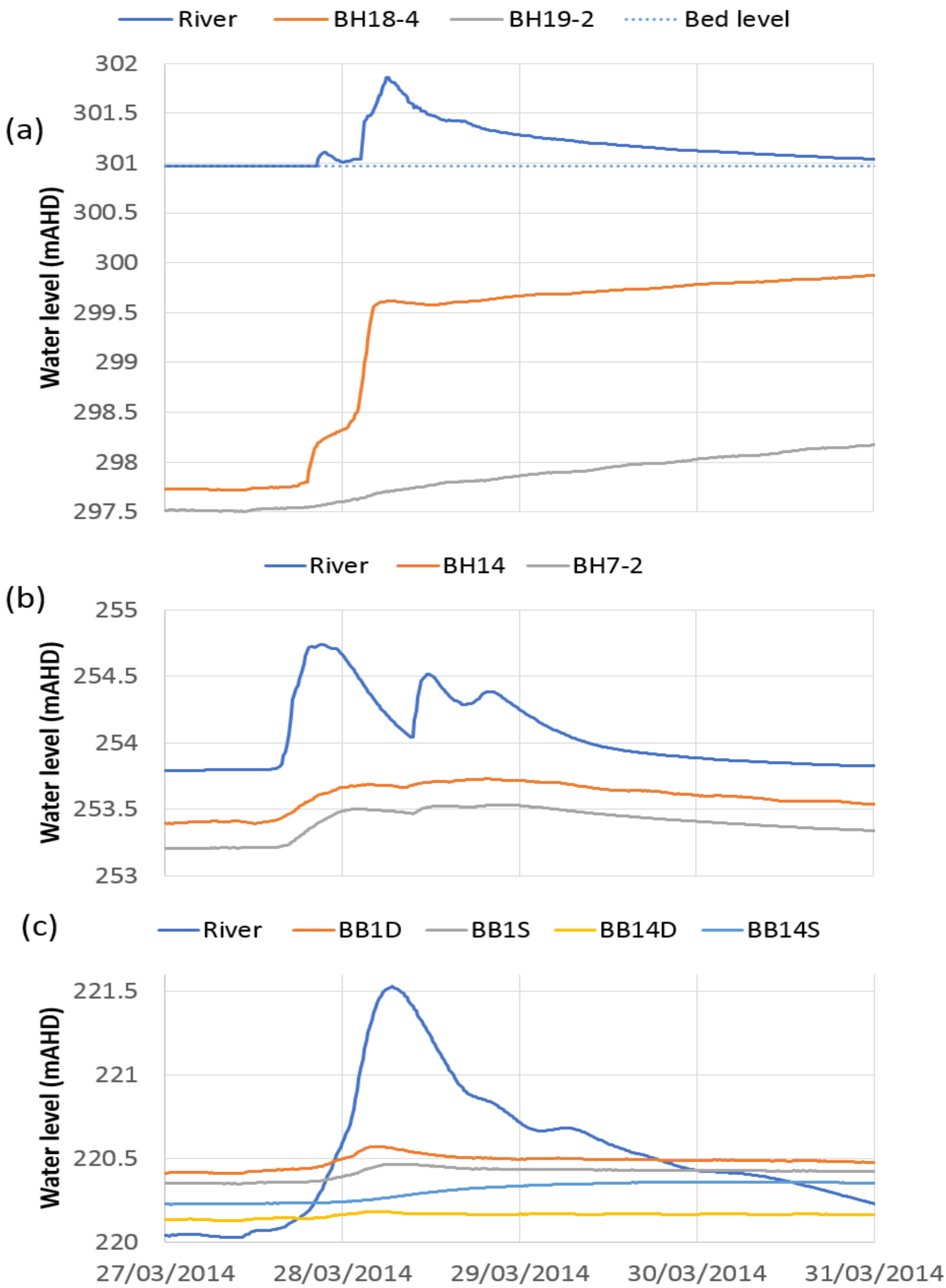
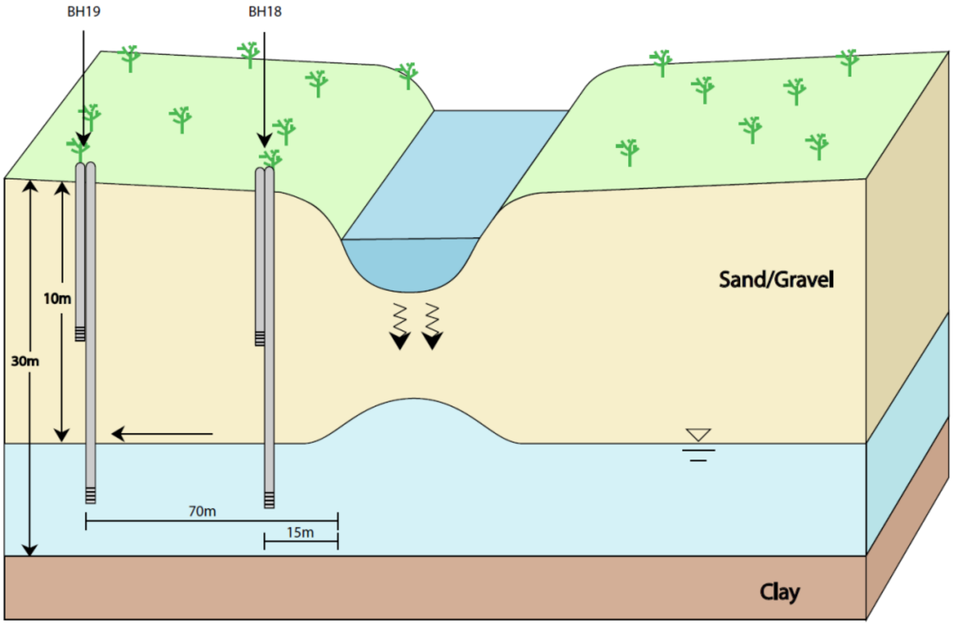


Figure 3.2 Three representative sites at (a) MC, (b) EC, and (c) BV along with their typical groundwater response to streamflow events. Y axis indicates water level in AHD. Note: due to the location of the stream level instrumentation, the measurements commenced from when the stream level reached 301 m AHD. It appears that the groundwater level of BH18-4 shown in (a) rises earlier than the river level. This is not the case.

3.1 Ephemeral section (Middle Creek Farm)

3.1.1 Site description

Figure 3.3 Installations (top) and schematics of the geological profile (bottom) at the site of Middle Creek Farm.



The disconnected section upstream in the catchment is represented by the monitoring bores at Middle Creek Farm (MC; Figure 3.3). Middle Creek is an intermittent stream classified as 4th order under the Strahler stream ordering system (Strahler, 1957). Most of the time it has no baseflow and only carries storm discharge from the Nandewar Ranges after heavy rainfall of approximately 100 mm in 7 days. Boreholes BH18 and BH19 were installed at 15 m and 70 m from the creek, respectively. Levels and depth of boreholes are provided in Table 3.1. Based on borehole lithological logs, the thickness of coarse sand and gravel exceeds 5 m at the site, which indicates storage capacity within an unconfined aquifer. Previous field observations indicated a clay layer under the landscape surface near the creek (Andersen and Acworth, 2009). Therefore, the stream channel serves as a preferential recharge path over diffuse recharge through the land surface. The streambed elevation is approximately 3 m deeper than the bank and streamflow is contained within banks for most flow events.

3.1.2 Groundwater level response to March 2014 storm event

Examples of typical groundwater level responses to high streamflow events are illustrated in Figure 3.2 using a storm event occurring in March 2014. Prior to the high streamflow event in Figure 3.2a, groundwater levels were between 297.5 m and 298.0 m, which leaves an unsaturated zone of approximately 3.2 m thickness under the creek bed. Limited by the location of the stream water level monitoring device as mentioned in Section 2.6, at least 3 hours of initial creek level variations were missing from the stream hydrograph. Therefore, the groundwater level of BH18-4 in Figure 3.2a appears to rise earlier than the river level, which is not the case. Despite the missing initial stream levels, the stream and BH18 hydrographs indicate a good alignment of the initial rises in water level. As the stream stage increased, a downward vertical hydraulic gradient was maintained towards the aquifer and raised the groundwater level. The initial 0.1-m stream peak caused a 0.5-m increase in the groundwater level. Following the maximum stream stage rise of 0.9 m, the groundwater level increased by 1.5 m. For each peak in the groundwater hydrograph, the highest rates of water level rise were recorded less than 2 hours after the initial time of groundwater response. Following the peak flows, the increasing trend of the groundwater level remained for 5 days with a rate of increase in excess of 0.05 m/day and remained steady until flow in the stream ceased completely.

Site	Borehole	Ground Elevation (mAHD)	Screen depth (m from ground)	Approximate Screen Elevation (mAHD)
MC	BH18-4	303.26	22-23	281
MC	BH19-2	302.75	21-22	281
EC	BH7-2	258.2	18	240
EC	BH12	258.4	11	247
BV	BB1-S	226.5	8.5-9.5	217.5
BV	BB1-D	226.5	15-16	211
BV	BB14-S	225.5	7.3-8.3	218
BV	BB14-D	225.5	15-16	210

Table 3.1 Details of monitoring boreholes.

At BH19, which is 70 m away from the creek, the hydrograph pattern differed from that recorded next to the stream. Although groundwater level response commenced at a similar time as for BH18, no clear peak was observed during the high stage event. After the peak, however, the rate of groundwater level increase was approximately 0.04 m/day higher than that in BH18 until the surface flow had ceased.

3.1.3 Summary of events

Changes in groundwater level during streamflow events are summarised in Table 3.2. An increase in groundwater level within the first 24 hours of each event showed variation between runoff events and between boreholes. The highest increase was recorded at BH18 in Jun 2016, with a 1.9-m increment. The water level variation was shown by three indicators:

(1) changes from pre-event water level to the highest value in the first 24 hours;

(2) the maximum value of the peak during the event; and

(3) the magnitude of the increment.

Means and standard deviations were also calculated for these descriptive values. For a total of seven events recorded between July 2012 and August 2017, stream level data were available for three events. The mean peak groundwater level was 300.6 m at BH18 and 299.8 m at BH19, with a standard deviation of approximately 0.4 m.

Table 3.2 Summary of events observed at MC.

	24-hrs increment (m)		Peak level (m AHD)		Peak magnitude	
	BH18-4	BH19-2	BH18-4	BH19-2	BH18-4	BH19-2
Jul-12	0.66	---------	---------	300.68	1.14	--------
Jan-13	1.79	---------	300.25	300.75	2.49	1.99
Jun-13	0.64	---------	---------	300.13	0.69	--------
Mar-14	1.82	0.27	299.61	300.29	2.49	2.06
Aug-14	1.28	0.14	299.59	300.34	1.73	1.29
Jun-15	1.74	0.18	299.43	300.57	3.31	2.42
Jun-16	1.92	0.25	300.27	301.28	4.4	3.63
Mean	1.41	0.21	299.83	300.58	2.32	2.28
St Dev	0.56	0.06	0.4	0.38	1.28	0.86

Increases in groundwater level within the first 24 hours of each event showed variation between runoff events and between boreholes. The highest increase was recorded at BH18 in June 2016 with a 1.9-m increment. By contrast, increases in BH19 were rather gradual, with a mean of 0.21 m. For each of the streamflow events, the groundwater level increase in BH18 was only about 20% higher than that in BH19. A record of observed events indicated that, when the initial groundwater level is low, the increase of water levels in the first 24 hours and during the entire event is high. The number of observed events for each site is less than 10, which is a relatively small sample size and not suitable for statistical analysis using methods such as Pearson’s correlation analysis. Consequently, a qualitative statement was made based on observation of events. The records of observed events indicated that, when the initial groundwater level is low, the increase in groundwater level in the first 24 hours and during the entire event is high.

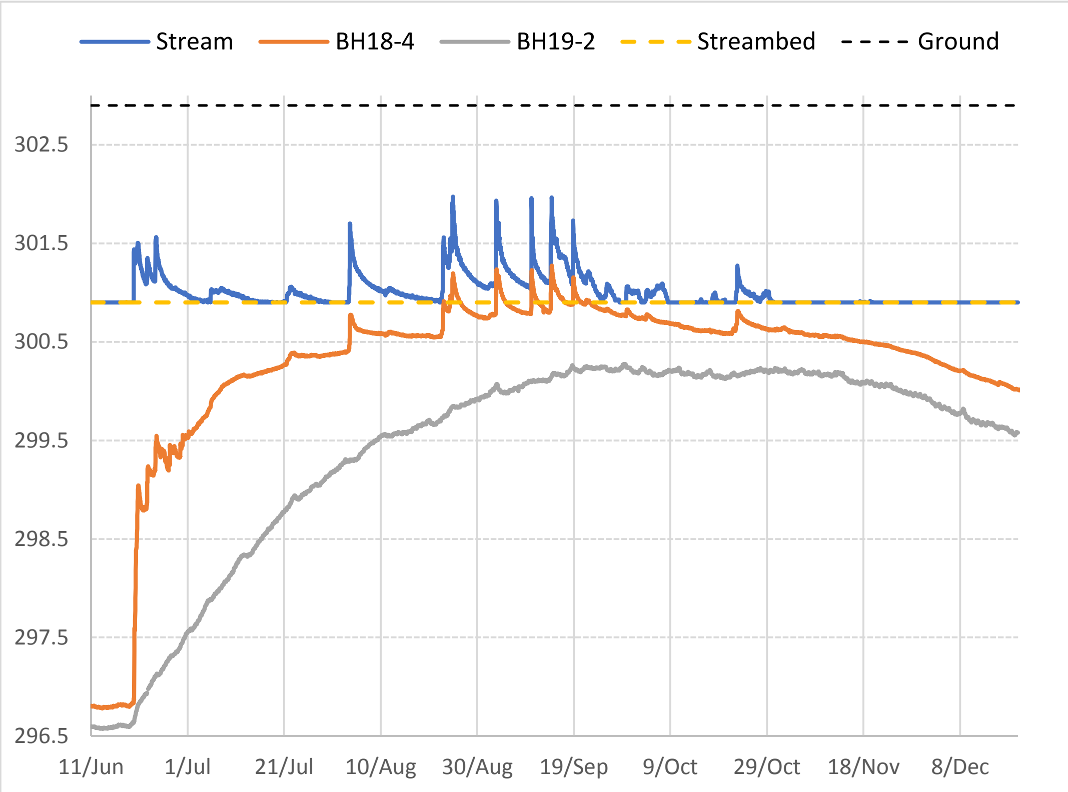


Figure 3.4 Groundwater level and stream stage at Middle Creek Farm (MC) observed from June to December 2016. It illustrates change from disconnected to connected conditions.

For extended dry or wet periods, the mode of surface water-groundwater interactions may change (for the definition of dry/wet period see section 4.2.1). Figure 3.4 shows an example of switching between the connected and disconnected mode during a wet period of 4 months following an 8-month dry period in 2016. At the disconnected reach of MC, the thickness of the unsaturated zone between creek bed and the groundwater table prior to the wet period was approximately 4 m. With the infiltration from the first flood in late June, the groundwater level at BH18 rose by 2.1 m. A series of subsequent flow events had maintained flow in the creek. Consequently, infiltration raised groundwater level and reduced the thickness of the unsaturated zone. By late August, the groundwater level was close to creek level. In September, no further increase in groundwater level was observed following subsequent floods of a similar magnitude as the first event. This latter part of the record in Figure 3.4 behaved similarly to the hydraulic pattern observed at EC and indicated a connected condition with a dynamic equilibrium. Further from the creek at BH19, the groundwater level continued to increase from mid-June to late September. However, as the lateral groundwater flow from the creek to BH19 is slower than the vertical infiltration from the creek to BH18, a time lag between BH18 and BH19 is observed. The observed level at BH19 was still approximately 0.5 m lower than that of BH18, providing evidence for a groundwater mound under the stream and continued lateral groundwater flow despite the dynamic equilibrium in the groundwater table near the creek. Additionally, as the horizontal hydraulic gradient reduced, the rate of increase in BH19 also diminished and began to drop when streamflow ceased in November. Eventually the system again became disconnected again as the flow events ceased.

3.2 Perennial section (Elfin Crossing)

3.2.1 Site description

The site of Elfin Crossing (EC) is next to the NSW state Avoca East (AE) gauging station. At this site, discharging groundwater sustains perennial conditions (Andersen and Acworth, 2009). Along the reach approximately 2 km upstream of EC, field investigations identified zones of strong groundwater upwelling in the form of springs and diffuse groundwater seepage at the upstream edge of a pool-riffle system. Downstream of EC, the stream again loses water into the underlying floodplain alluvium. The distance of dry conditions between the last permanent pools and the confluence with the Namoi River varied between 6–10 km depending on the preceding rainfall conditions. Only during and after severe storms, high streamflow from upper Maules Creek and its tributaries reaches past EC and provides temporary flow to the Namoi River.

Under normal conditions (i.e., without drought or storm), the groundwater level is generally similar to the streambed level, and the groundwater levels are in a dynamic equilibrium with the stream. The length of the gaining section varied with the longer-term climate. During fieldwork, a shorter length was observed during extended dry periods and a longer length was observed in wet periods.

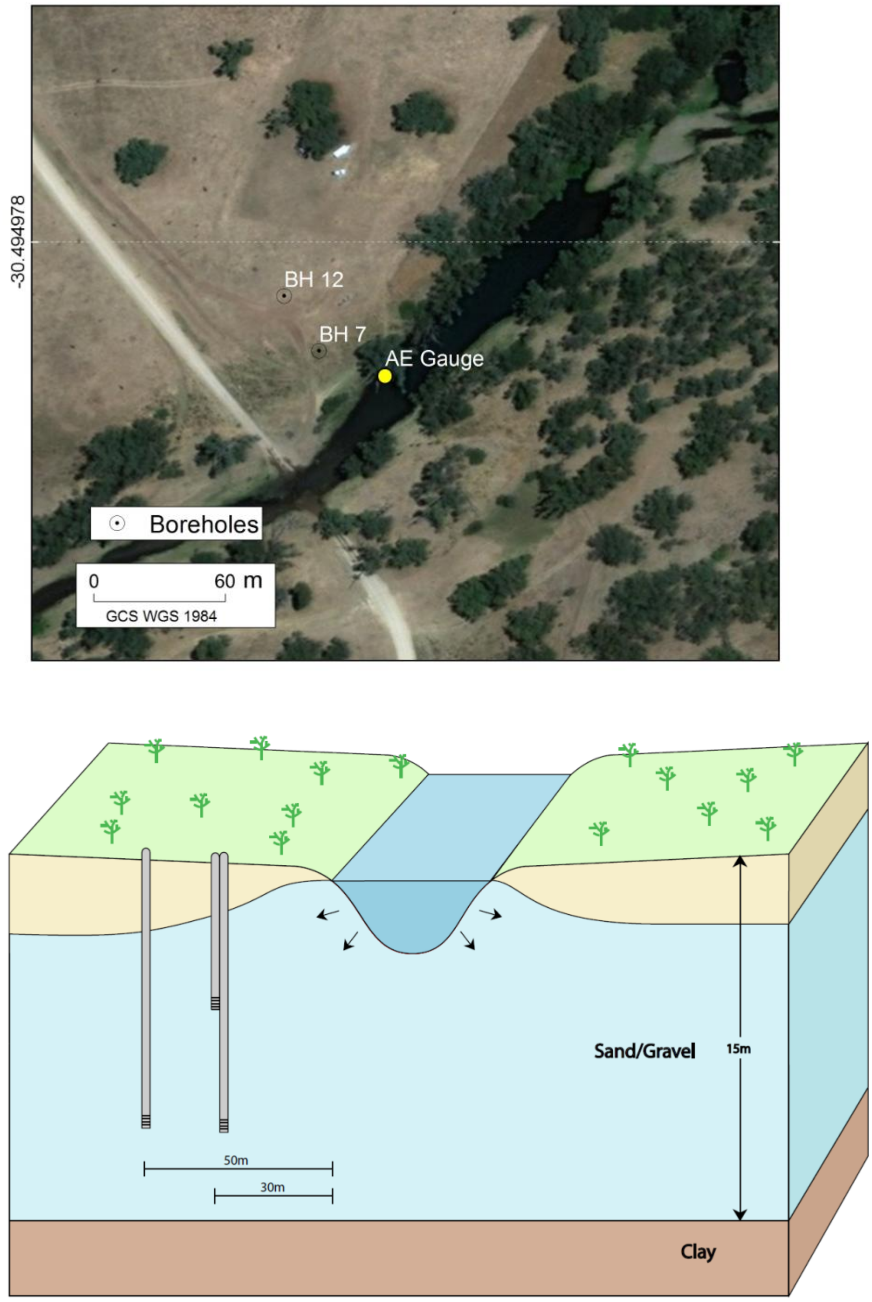


Figure 3.5 Installations (top) and schematics of the geological profile (bottom) at the site of Elfin Crossing.

3.2.2 Groundwater level response to the March 2014 storm event

Prior to the storm event, the groundwater level in BH7 at EC was 0.4–0.6 m lower than the stream level (Figure 3.2[b]). Groundwater levels, measured in BH12, were at 253.4 m, which was similar to the streambed elevation. Three stream peaks were observed during the March 2014 storm. The first stream peak of 0.9 m raised the groundwater level by approximately 0.3 m. As the peak receded, groundwater levels immediately stopped increasing and started receding. While the second peak was smaller in both relative and absolute magnitude, it caused a further 0.1-m increase in the groundwater levels. Unlike the sustained increase at MC, the groundwater level at EC dropped at both boreholes once the creek water level started to recede. Changes in the groundwater table from the pre-flood level decreased to less than 0.1 m within 3 days after the event peaked and completely disappeared within 7 days.

3.3 Bellevue Farm

3.3.1 Site description

Bellevue Farm (BV) is located on the eastern bank of the Namoi River on the floodplain (Figure 3.6). The average site elevation is 227 m AHD, which is approximately 8 m above the channel bottom. Fifteen nested boreholes on the east of the river formed a 150-m transect perpendicular to the river. Piezometers were installed in pairs, with a shallow (~10 m below ground level) and a deeper (~16 m below ground level) piezometer. A farm irrigation bore was located at the end of the transect furthest from the river, which supplies water for furrow irrigation. The site is layered with multiple shallow aquifers. The top is unconfined with topsoil and silt at the surface. A low permeability layer of clay and sandy clay separate the deep semi-confined aquifer.

3.3.2 Groundwater level response to March 2014 storm event

Although Maules Creek may provide a significant contribution to the Namoi River during flood, most high stream level events at BV are the results of rainfall and runoffs in the regulated Upper Namoi Catchment (Andersen and Acworth, 2009). Prior to the March 2014 event, the groundwater level at BV was above the streambed level (Figure 3.2[c]). Few changes in the groundwater hydrograph were identified until the rising limb of the stream hydrograph exceeded the groundwater level at midnight on the 28th of March. A 1.5-m stream peak occurred at 6:30 a.m. Rises of 0.16 m and 0.11 m were observed from BB1D and BB1S, respectively, and dissipated along with the recession of the stream peak. However, the gains did not dissipate completely. Each borehole gained 0.09 m during the event and receded at a rate of 0.01 m/day.

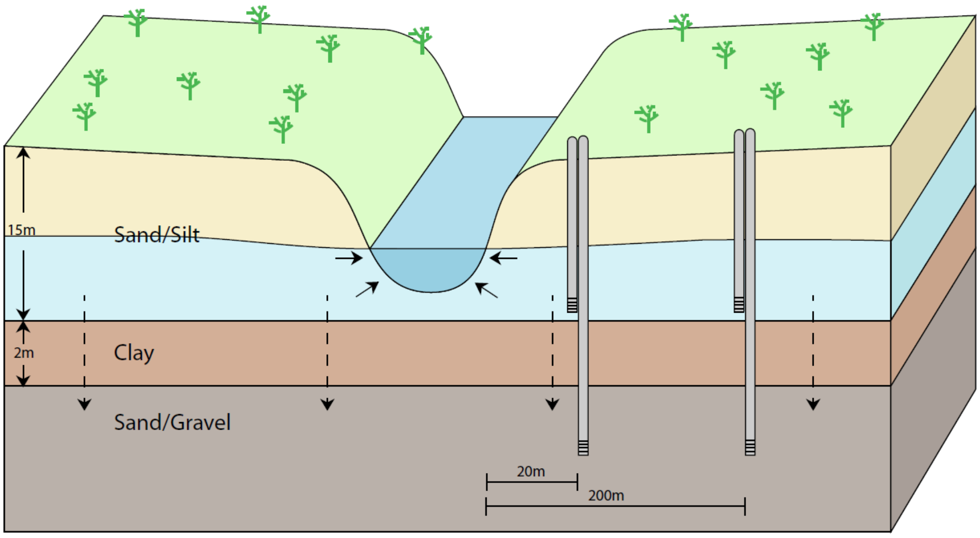


Figure 3.6 Installations (top) and schematics of the geological profile (bottom) at the site of Bellevue Farm.

3.3.3 Intra-annual variation

Groundwater levels at BV throughout 2015 are presented in Figure 3.7, with pumping drawdown and high stream stage events driving fluctuations. Groundwater levels were generally higher than the river level, except during high streamflow and pumping events. It is noted that the shallow bores BB1S and BB14S varied in a similar pattern. On the contrary, the level of BB14D dropped during the pumping season and gradually recovered from March to September.

3.4 Discussions

3.4.1 Groundwater recharge volumes depend on antecedent conditions

Along intermittent stream reaches, the flux of water across the interface between surface water and groundwater only occurs in a downward direction during an episodic flow event. This provides important recharge input in arid and semi-arid environments, yet the resulting recharge volume is highly variable between flow events (Harrington et al., 2002, Scanlon et al., 2006). This study illustrated that the initial groundwater level in the underlying aquifer is a significant factor controlling the amount of recharge relative to the size of the discharge event.

Neither rainfall nor stream discharge occurred with a clear seasonal pattern in this semi-arid environment. Although the rainfall record shows a seasonal pattern on the decadal scale (Figure 4.5), rainfall occurrences within each individual year vary. The standard deviation of the monthly rainfall on the alluvial plain is 49.2 mm, which is almost equivalent to the mean of 50.5 mm (Figure 3.1). Between 2014 and 2017, the running 14-days cumulative rainfall exceeded 100 mm on seven occasions. Four such rainfall events led to streamflow in Maules Creek. On the other hand, annual temperature variations and, hence, the natural water loss in the form of evapotranspiration shows a clear periodic pattern of seasonality. This climate condition means a steady natural water demand was placed over a highly variable supply. Therefore, if a high streamflow event occurs at a different time, it may flow onto a dry or wet aquifer, which will lead to a different amount of groundwater recharge. The duration since the last wetting event is critical to the recharge, which is also irregular. The time between these events, between 2014 and 2017, varied from 28 to 516 days.

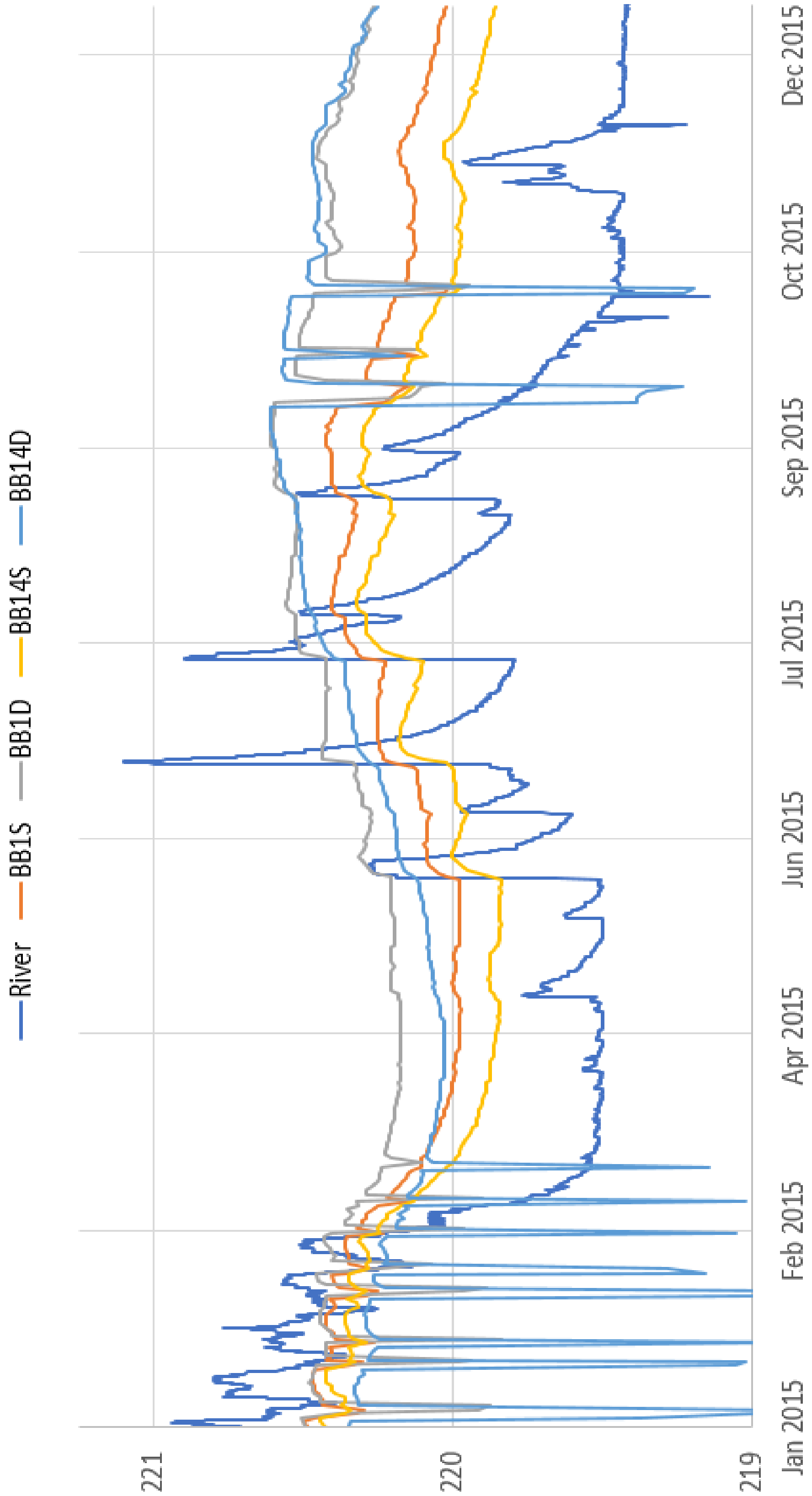


Figure 3.7 Groundwater fluctuation throughout a 12-month period at the BV site.

For a disconnected reach within a given geologic setting, the potential for accepting recharge largely depends on the thickness of the unsaturated aquifer and the groundwater elevation at the time of an event. During dry periods, the groundwater level beneath ephemeral streams undergoes a gradual recession, as described by Cuthbert et al. (2016). With a lower groundwater level, more space is available for accepting potential recharge. As the time between streamflow events increases, the groundwater level further decreases, allowing more storage for subsequent events. When a streamflow event occurs, infiltrating water tends to fill the unsaturated space. The streambed surface sets the upper limit of the unsaturated zone storage space. This explains why the highest water levels for each event were similar for events above a certain size, regardless of the duration or the scale of the streamflow event.

In contrast, along a connected reach of a perennial stream under natural conditions (i.e., for streams not heavily affected by groundwater abstraction), groundwater and surface water are in a dynamic equilibrium. This means that the groundwater table at the streambank is similar to the surface water and rapidly responds to changes in the surface water level. Such equilibrium is formed by the combination of precipitation and evapotranspiration across the catchment, determining the general groundwater level (further controlled by lithology and topography), as well as the streamflow characteristics (Fan et al., 2007). The lack of an unsaturated zone beneath the streambed and the high groundwater level below the banks means there is generally little unsaturated storage space left to accept any further recharge from the stream during periods of elevated streamflow. The main storage mechanism for such streams is bank storage, which is typically transient storage with a duration of weeks to months (Kondolf et al., 1987, Unland et al., 2015). Such equilibrium is commonly observed in many perennial streams in humid environments (e.g., Menció et al., 2014, Balbarini et al., 2017). Observations at EC indicated that this portion of the aquifer system is connected and behaves like a humid perennial stream. Consequently, along this reach of Maules Creek there is limited potential for groundwater recharge from the stream during high streamflow events. While a high stream stage event creates a hydraulic gradient towards the aquifer which would potentially allow for recharge, the available storage space in the aquifer is limited and is above the low-flow stream level. This means that, once the flow event recedes, the hydraulic gradient is reversed towards the stream and groundwater is discharged back into the stream. Any gains in groundwater level tend to re-equilibrate with the receding stream level and, therefore, largely become bank storage (Vázquez-Suñé et al., 2007). Based on observed hydrographs, the exchange of flux between stream and aquifer related to an event is only on the timescale of one to three weeks. This converges with the reported timescale for gravel/sand stream bank material simulated by Whiting and Pomeranets (1997).

Another gaining stream example is provided through observations from BV. At the lowest part of the study catchment, Namoi River naturally receives regional groundwater flow discharge into the stream as groundwater contributes to the baseflow. As shown in Figure 3.7, groundwater levels are higher than the low-flow stream level; hence, hydraulic gradients are towards the river. Limited by the low hydraulic conductivity layers between the river and the lower semi-confined aquifer, a hydraulic gradient is maintained towards the river and did not reach equilibrium as in EC. Despite the reversal of the hydraulic gradient observed during the drought in the late 2000s (McCallum et al., 2013, Kelly et al., 2013), this reach predominantly showed groundwater discharge during the period studied in this chapter. Although BB14D showed a trend of drawdown followed by gradual rising (Figure 3.7), its level remained higher than stream level for most of the year. During the high streamflow event, no sharp increase could be identified. The start and finish time of the rise suggested this is due to regional groundwater flow following pumping seasons.

Being a naturally gaining reach, the aquifer beneath the Namoi River alluvium has little available unsaturated space for accepting long-term groundwater recharge. Changes in the groundwater level in response to high streamflow events were small compared with that observed at MC. The semi-confined aquifer showed instantaneous variations when changes in stream level were observed. This may indicate a loading effect in the groundwater aquifer rather than direct recharge (van der Kamp and Schmidt, 2017). Alternatively, given the assumed heterogeneity of sediments, preferential flows may enter and exit the aquifer through zones of coarser materials. Both mechanisms only provide temporary recharge during and following the event.

3.4.2 Stream classification in semi-arid/arid environments requires consideration of the subsurface lithology and climate conditions

Representation of surface water-groundwater interaction is required in most hydrogeological studies and models with surface water features. This usually involves classifying the streams of interest (Barnett et al., 2012). With the dynamic nature of surface water-groundwater interaction highlighted by many studies (Covino and McGlynn, 2007, Banks et al., 2011, McCallum et al., 2013, Rau et al., 2017, Xian et al., 2017), it is now generally agreed that the gaining or losing conditions are not static in time for a given stream or reach.

However, streams in the semi-arid study area are not clearly defined under such a classification system. Due to the highly variable rainfall input, the flow regime of Maules Creek varies from dry streams to isolated pools and continuous flow in both time and space. A similar variable flow regime is also found in the Namoi River, as indicated in the flow duration curve in later sections and in Figure 4.8(a).

Observations from EC and BV show examples of a variable stream regime. In the perennial section of EC, although comparison between the stream stage and groundwater level indicate the stream is in a dynamic equilibrium with the groundwater table, field investigation by Andersen and Acworth (2009) near EC identified that the extent of the gaining reach varies with seasonality and climatic condition, with dryer conditions reducing the length of reach of visible groundwater discharge to the 1- to 2-km reach upstream of EC. Similarly, at BV, the stream condition varies between periods of high and low streamflow. For connected losing streams, the transmissivity or hydraulic conductivity also influences the recharge acceptance capacity. The deposition sequence provides zones of different transmissivity. More recharge is possible if the flood waters enter an aquifer unit with a higher transmissivity.

Lengthy events or several smaller events in rapid succession can also change the connection state of a stream reach. As the example of the disconnected reach of MC has shown, (section 3.1.3 and Figure 3.4), the 4-m thick unsaturated zone became saturated following a wet period of 4 months. During this period, the stream was equilibrated with the groundwater level, which is similar to perennial behaviour. This reach effectively became connected and any future high streamflow events can cause the limited increase in the groundwater level. As the surface flow ceased, the groundwater level declined and the same reaches became losing connected, and, ultimately, losing disconnected. Using stream classification scenarios for describing a stream reach has its limitations, especially when the descriptions are based on point measurements (Menció et al., 2014, Wang et al., 2017).

4 Catchment scale analysis of groundwater drawdown recovery in natural climate variation

4.1 Overview

Natural periodic climate variability affects the catchment water balance through variable yearly precipitation, evapotranspiration, stream runoff, and the recharge and discharge of groundwater in the alluvial aquifer (Muttiah and Wurbs, 2002). In alluvial plains that have been developed for agricultural production, changes in irrigation demands and, thus, the groundwater abstraction rate add more uncertainty (Chen et al., 2010). In semi-arid environments, where surface water resources can be particularly erratic and unreliable, groundwater offers a buffer through drought periods caused by natural climatic variations (Tsur and Graham-Tomasi, 1991, Acworth et al., 2015). While groundwater abstraction during drier periods could be compensated by natural recharge during wetter periods, it is difficult to estimate and quantify the distribution and magnitude of recharge under different dry/wet conditions (Ajami et al., 2015). Therefore, the effective management of groundwater allocations should not solely rely on static estimations of the recharge rate; it needs to take into consideration the spatial and temporal variability in recharge (Shamsudduha et al., 2012, Ghosh et al., 2014).

This portion of the study focuses on assessing the groundwater recovery during multiple changes between wet and dry periods using patterns identified in the previous chapter at the catchment scale. The groundwater dataset is spatially distributed over the semi-arid catchment from the mountain front to the riverine area, and temporally spans ten years before and twenty years after the onset of major groundwater abstraction. It focuses on distinguishing the links between groundwater level fluctuations and climatic variation. It also determines the dominating groundwater recharge mechanisms, particularly the role of recharge through a high stream stage event in recovering groundwater drawdown.

4.2 Data source and method of analyses

4.2.1 Precipitation and dry/wet period definition

Daily precipitation data were obtained from the two nearest Bureau of Meteorology (BOM) weather stations (MTL: Mt Lindsay 055058; TUR: Turrawan 055041; refer to Figure 2.1 for locations). Data patterns of discrete daily rainfall measurements and gradual variation in groundwater levels are quite different. Between the available moisture above ground and the storage in an aquifer, the unsaturated zone acts mathematically as a transfer function that filters the signal between rainfall input and groundwater level response (Mattern and Vanclooster, 2010). The cumulative rainfall departure (CRD) curve was used to represent the overall changes in water availability and the long-term climate trend. The CRD was formed by summing daily observations of rainfall into monthly steps, and subtracting the mean monthly rainfall of the dataset from each time step to obtain the departure. The residual departure values are subsequently accumulated. For time step j,

$$CRD_j = \sum_{i=1}^{j}(x_i - \bar{x})$$

where x_i is the rainfall summed for time step i (Xu and Van Tonder, 2001, Baalousha, 2005). This method is commonly used in visualising for visualizing the variability of precipitation against the average of the record and in identifying periods of abnormalities embedded in a time series (Weber and Stewart, 2004). The downward slope in the CRD curve shows periods of below-average precipitation while upward slopes show periods of above-average precipitation (Figure 4.6).

The Southern Oscillation Index (SOI) was also obtained from the BOM to represent the variability in the regional climate. The SOI observed atmospheric pressure differences between Darwin and Tahiti, which have been used as an indication of temperature change on the central and eastern tropical Pacific Ocean. Sustained negative values of the SOI indicated the occurrence of El Niño events (Hanley et al., 2003).

The dry and wet periods were determined using the rainfall deciles method, which was developed and adopted by the Bureau of Meteorology and is currently used in defining

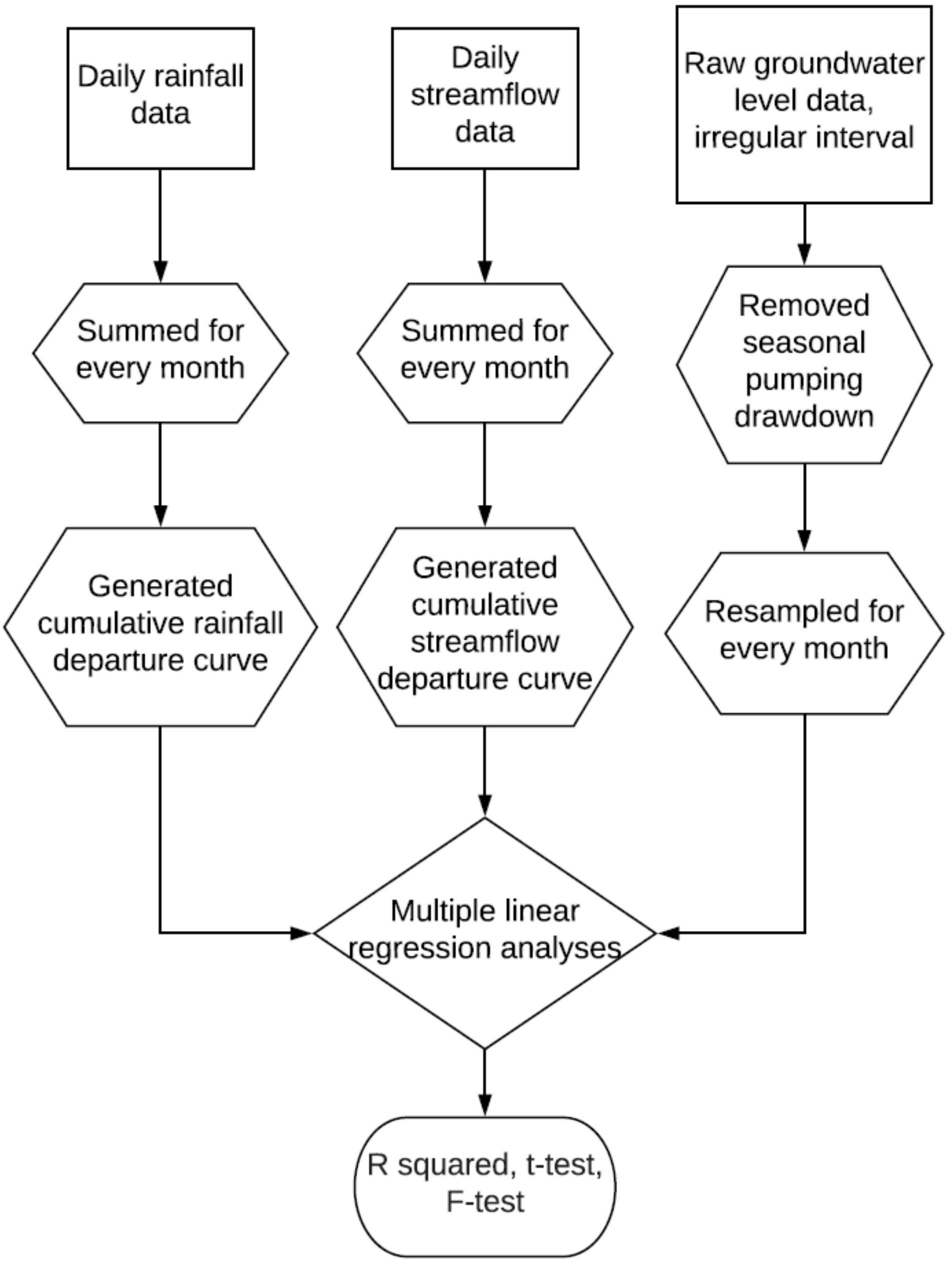


Figure 4.1 Steps of data analysis.

drought and wet periods in Australia (Gibbs and Maher, 1967). A drought, or serious rainfall deficiency, is defined as the observed 3-month rainfall lying below in the lowest 10% of the long-term record; whereas, the lowest 5% is referred to as a severe rainfall deficiency. The 30th percentile is used as the threshold of below-average rainfall. To clear a meteorological drought, rainfall needs to be above the 30th percentile over a 3-month period (Wilhite and Glantz, 1985). Similarly, the highest 5th, 10th and 30th percentiles are used as the threshold of severe, serious and above-average rainfall surplus conditions, respectively. The 3-month minimum threshold ensures that the impact of shorter events is minimised (Wilhite and Glantz, 1985).

4.2.2 Stream discharge

Stream levels and discharges were obtained from stream gauges managed by the water resource management authority in New South Wales, currently known as the Department of Industry—Water. Of the three gauges in the catchment, two are on the Namoi River (BOG: Boggabri 419012; TUR: Turrawan 419023; Figure 2.1). Stream water levels have been measured at these locations since the mid-1970s. Stream discharge rates and volumes were derived with the use calibrated rating curves (NSW DI Water, 2015). Based on the uncertainty assessment of streamflow rating curves (Tomkins, 2014), the gauging data is regarded as good quality, with 89.6% and 95.0% of the time at BOG and AE, respectively. Therefore, the streamflow analysis in this study was based on the derived flow rate.

The streamflow characteristics of the Namoi River and Maules Creek were described with flow duration curves (Vogel and Fennessey, 1994). Since streamflow was featured with episodic high stream stage or flood peaks with low flows, its behaviours are quite different from the slow responses in the groundwater system. As a result, the stream discharge data was also converted to the monthly cumulative streamflow departure (CFD) curve using the same method as for cumulative rainfall departure (Blakers et al., 2011).

4.2.3 Groundwater hydrograph analyses

The data from the groundwater level has been manually dipped from 61 piezometers at 35 monitoring boreholes since the mid-1970s. The record had irregular measurement intervals, which varied between one to two months (NOW, 2015). Hydrographs of each piezometer were plotted for initial assessment using visual comparison. Example hydrographs were plotted on a catchment map (Figure 4.9). For simplicity of presentation, similar hydrographs from boreholes that are less than 2 km apart are not shown. Similar to McCallum et al. (2013), an upper and a lower piezometer were selected to plot the data for boreholes with multiple piezometers.

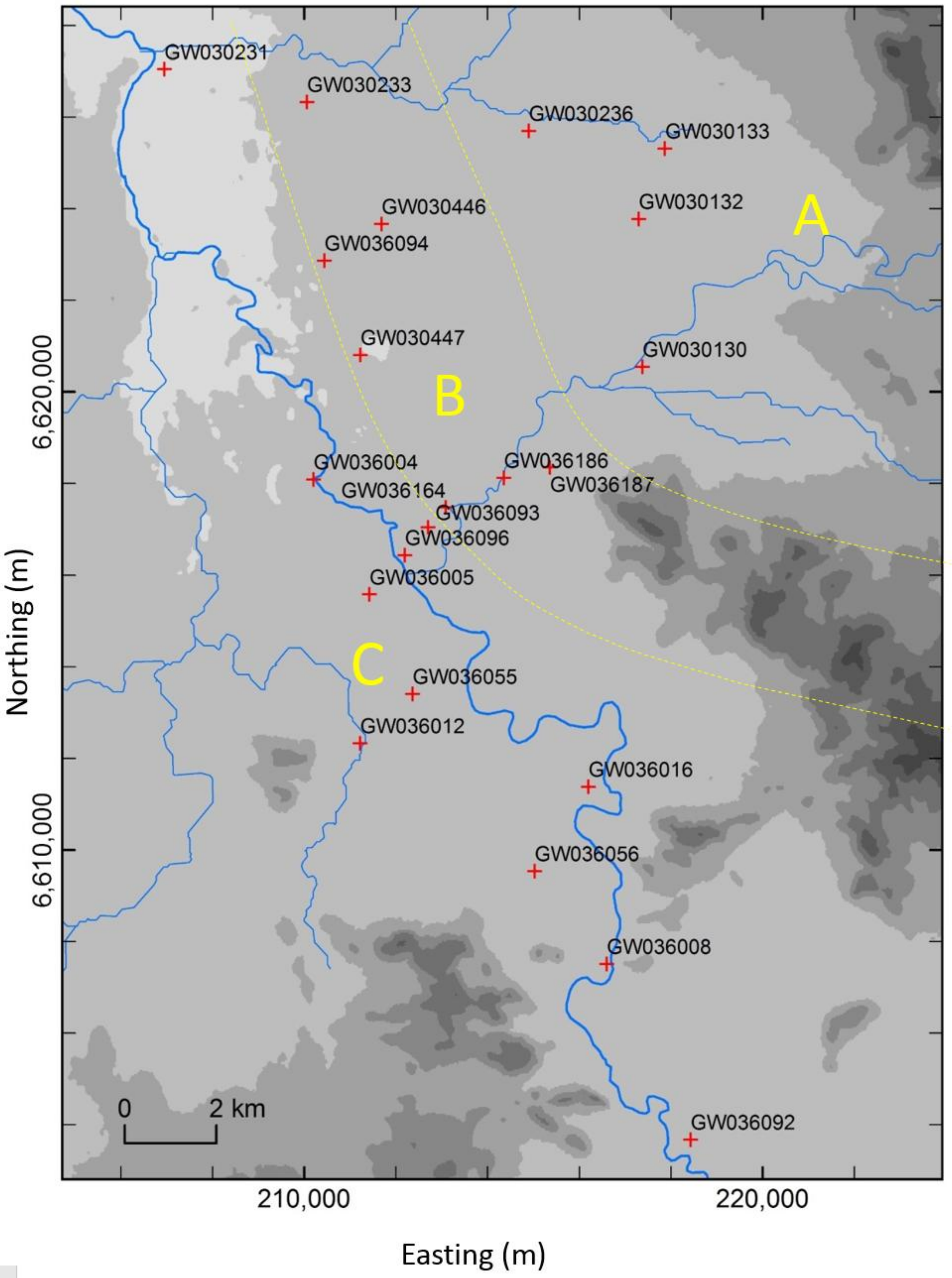


Figure 4.2 Locations of selected long-term groundwater monitoring boreholes with zones defined by long-term patterns of groundwater level.

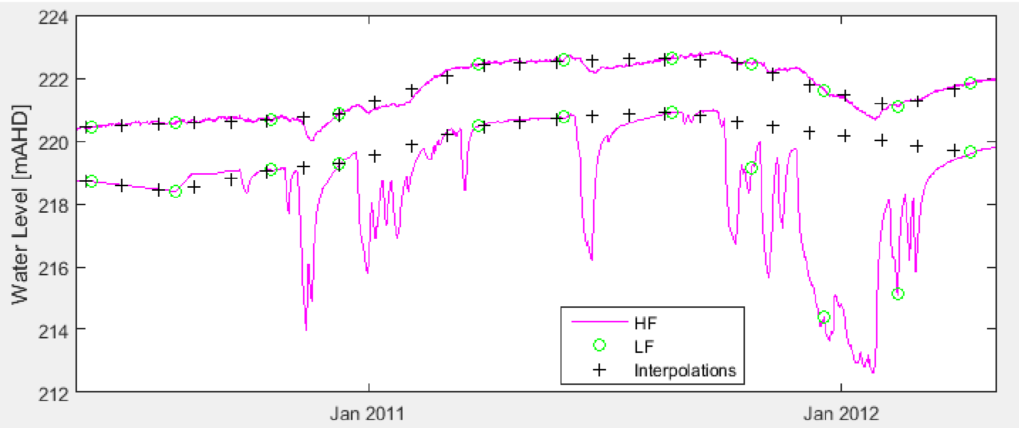


Figure 4.3 Validation of filtered and resampled groundwater data. **HF**: high frequency groundwater level data measured using automated pressure transducers with sub-hourly steps. Such data has only been available at a few boreholes since the late 2000s. **LF**: low frequency manually dipped data with irregular measurement intervals. **Interpolations**: resampled time series at monthly intervals with the pumping drawdown filtered.

The initial visual assessment identified three generalised categories of hydrograph shapes. Based on the initial findings, the catchment was subsequently divided into three zones with considerations for the geographic location of each borehole (Figure 4.2). Zone A boreholes are in the east of the catchment near the Nandewar Range with ground surface elevations ranging between 240 and 255 m AHD. Hydrographs in Zone A varied mainly over decadal scales. Boreholes on the alluvial plain between the river and the mountain front were categorised as Zone B. Hydrographs from these boreholes showed strong short-term changes on seasonal scales. Zone C includes boreholes within 2 km to the east and boreholes to the west of the Namoi River. Although some hydrographs in Zone C also showed seasonal changes, on the decadal scale, all hydrographs demonstrated overall declining trends (e.g. 30231 in Figure 4.9).

Many hydrographs included abrupt seasonal decreases, especially in the lower piezometers of the Zone B and C boreholes. The decrease provided evidence of the impact of groundwater abstractions during pumping seasons. At the start of each pumping season, hydrographs showed temporal drawdown responses. A rebound in water levels was also observed at the end of each pumping season. When doing long-term statistical analysis, however, the seasonal pumping drawdowns needed to be filtered out to avoid skewed datasets. The automated process of pumping drawdown removal from groundwater hydrographs was based on the comparison of two adjacent datapoints in the time series. A point was filtered out if

1. the slope between two points is lower than -0.016 m/day or higher than 0.055 m/day;
2. the following data point is 1.3 m lower than the current point; or
3. the difference between the current data point and the 5th preceding data point is greater than 5.

Note these criteria were identified empirically based on a visual inspection. This method of filtering was also validated at boreholes with available high-frequency (15 or 30 minutes) water level data (Figure 4.3). Data gaps after filtering out the observed groundwater levels during pumping were filled with linearly interpolated data points between the pre-pumping head and the recovered heads at the end of the pumping season. The filtered time series was then linearly resampled into monthly time steps, which was consistent with rainfall and streamflow data.

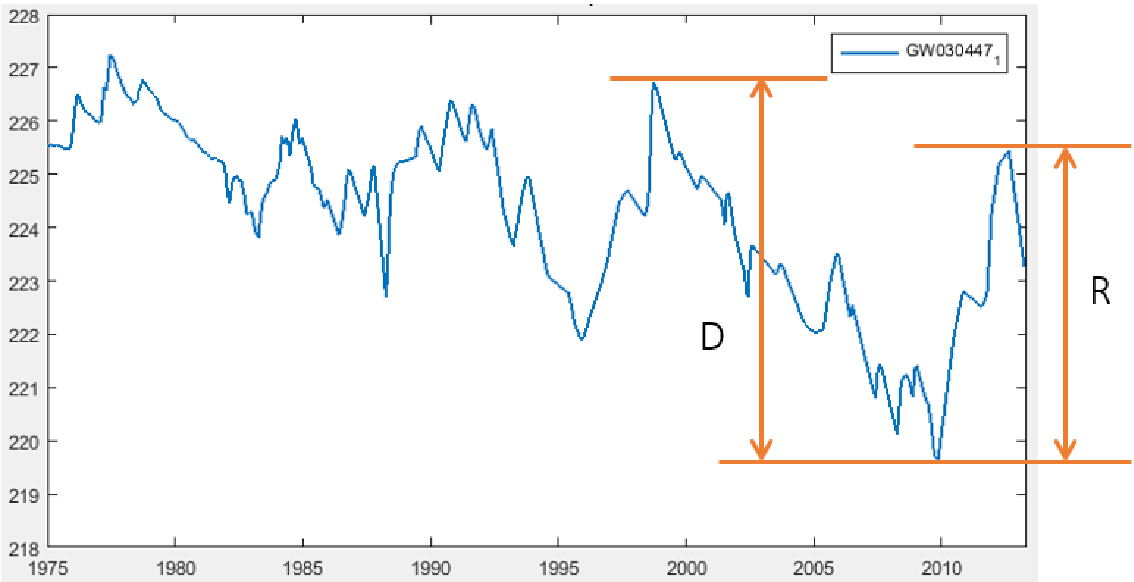


Figure 4.4 The concept of recovery ratio demonstrated on groundwater hydrograph of borehole GW030447-1.

4.2.4 Recovery ratio

Filtered groundwater levels at the beginning and end of each dry period were compared across the catchment at each monitoring location. To quantitively assess the extent of groundwater level recovery at the end of each dry period, the concept of *Recovery Ratio* is introduced as

$$Recovery\ Ratio = \frac{R\ (Recovery\ in\ wet\ period)}{D\ (Drawdown\ in\ dry\ period)}$$

The recovery ratio for each piezometer was automatically generated for each dry-wet transition period and is graphically represented in Figure 4.4.

4.2.5 Regression analyses

The strength of coupling between the groundwater level and environmental factors were analysed with multiple linear regression models. This method was used to explain the relationship between explanatory variables and a response variable by fitting a linear equation to the observed data. The population regression line for a total of n explanatory variables $x_1, x_2 ... x_n$ was defined as

$$y = \beta_0 + \beta_1 x_1 + \beta_2 x_2 + ... + \beta_n x_n + \varepsilon$$

where ε is the residual term and $\beta_0, \beta_1, \beta_2 ... \beta_n$ are the regression coefficients of each explanatory variable (Freedman, 2009). This calculation was carried in R using *The R Stats* package (R Core Team, 2013). In addition to calculating coefficients for each variable, the package provided three assessments for the statistical significance. A t-test showed the significance of each explanatory variable to y and, therefore, was used for model optimisation. Only explanatory variables with $p < 0.001$ were kept in the model. An F-test assessed the overall performance of the model, and r^2 showed how close the data are to the fitted regression line. Although the explanatory variables were not strictly independent (e.g., the dry/wet cycle of climate variability affects both rainfall in the catchment and streamflow upstream), it is understood that the frequencies of variation for each explanatory variable were quite different and, therefore, also the way they act on groundwater levels. Thus, influences from each factor can be revealed through the regression analysis. Data used for regression analyses are provided in the appendix.

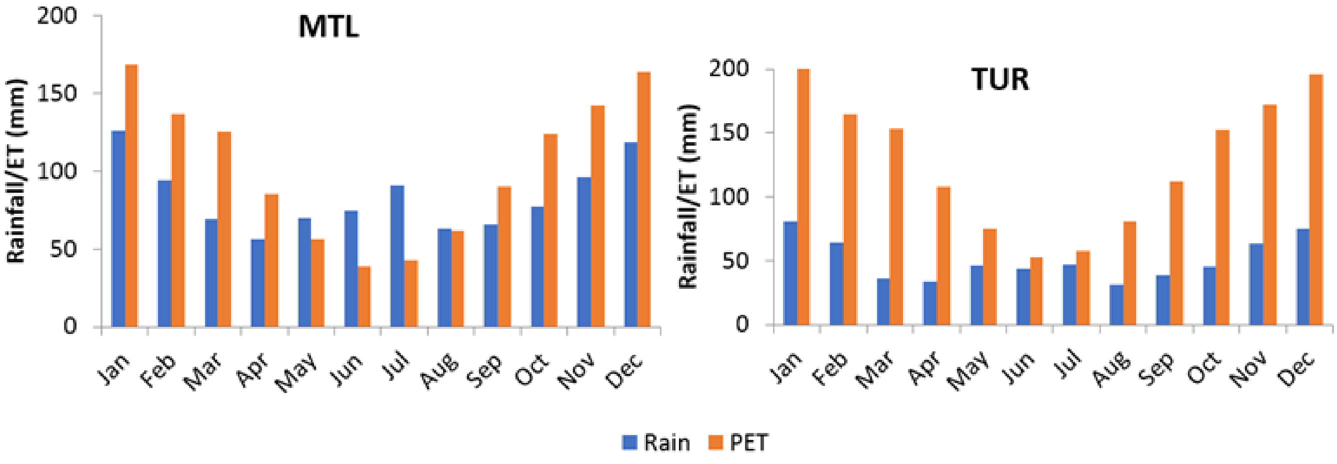


Figure 4.5 Monthly average climate information for MTL and TUR.

4.3 Results

4.3.1 Precipitation and dry/wet periods

The climate within the mountain range (as represented by the climate station MTL) is characterised by higher precipitation and lower potential evapotranspiration compared with the floodplain (climate station TUR; McCallum et al., 2013). Annual mean precipitation recorded at MTL is 1001 mm/a (based on the long-term record of 1889–2015), which is significantly greater than the 606 mm/a rainfall recorded at TUR recorded during the same time period (Figure 4.5).

The potential evapotranspiration (PET) at MTL for the same period is 1237 mm/a, which is 18% lower than at TUR (1526 mm/a). Within a year, rainfall is distributed throughout each season, with slightly more rainfall in summer quarters (approximately 35% of the annual rainfall). The daily net water balance at the ground surface (estimated by subtracting PET from rainfall) is negative most of the time. For each season, average PET generally exceeds the corresponding average rainfall with the exception of the winter quarter at MTL (Figure 4.5).

The dry and wet periods determined by the rainfall decile method are indicated using the red and green bar, respectively (Figure 4.6), with the cumulative precipitation trend shown as the CRD curve. During prolonged dry periods in 1980–83, 1993–94, 2001–04 and 2006–08 the curve shows downward trends. At their peak, annual

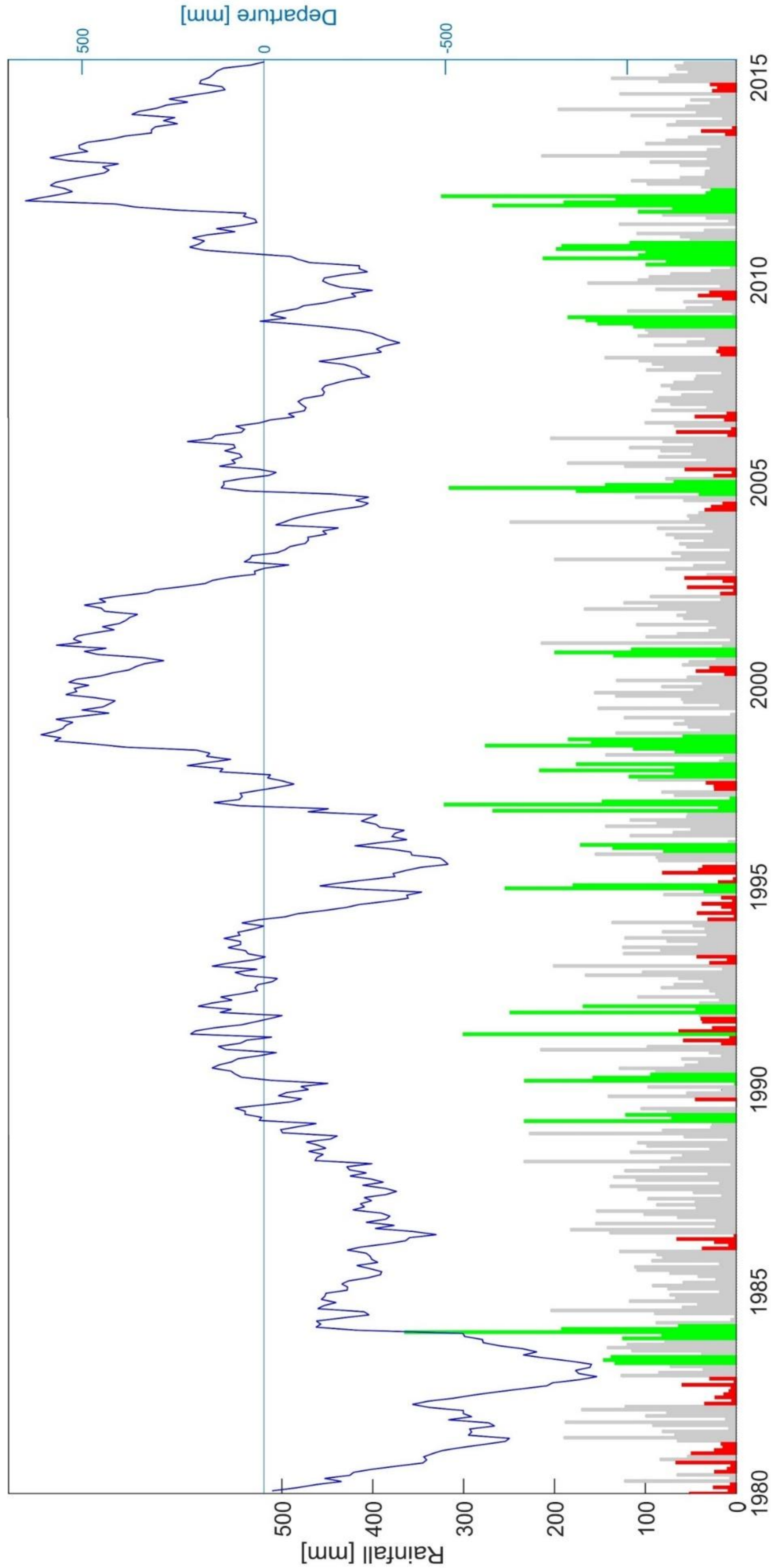


Figure 4.6 Monthly rainfall measured at TUR in the floodplain shown as the CRD curve between 1980 and 2015. Green and red bars indicate the wet and dry periods, respectively.

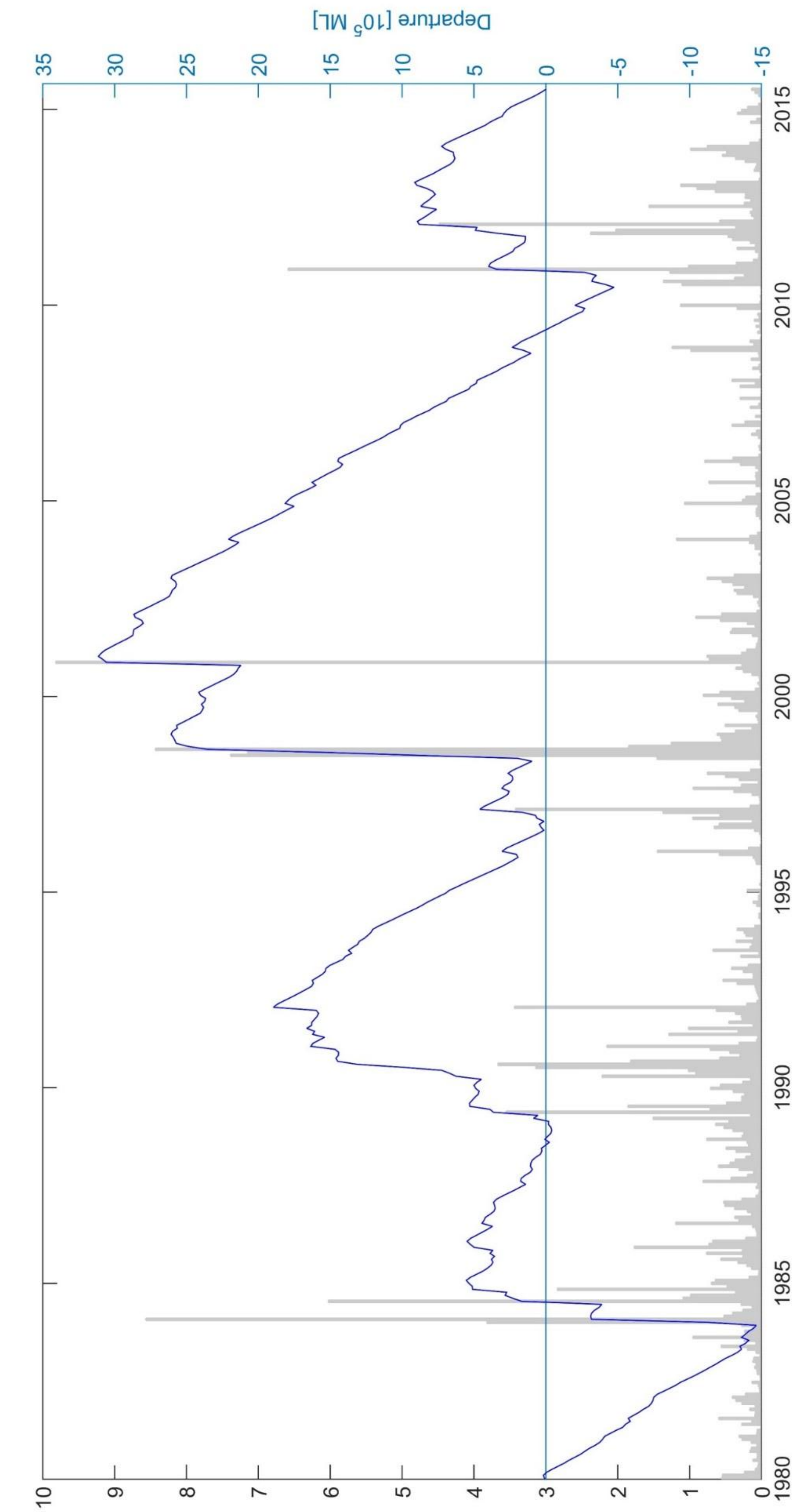


Figure 4.7 Streamflow of Namoi River with the CSD curve plotted on the secondary axis.

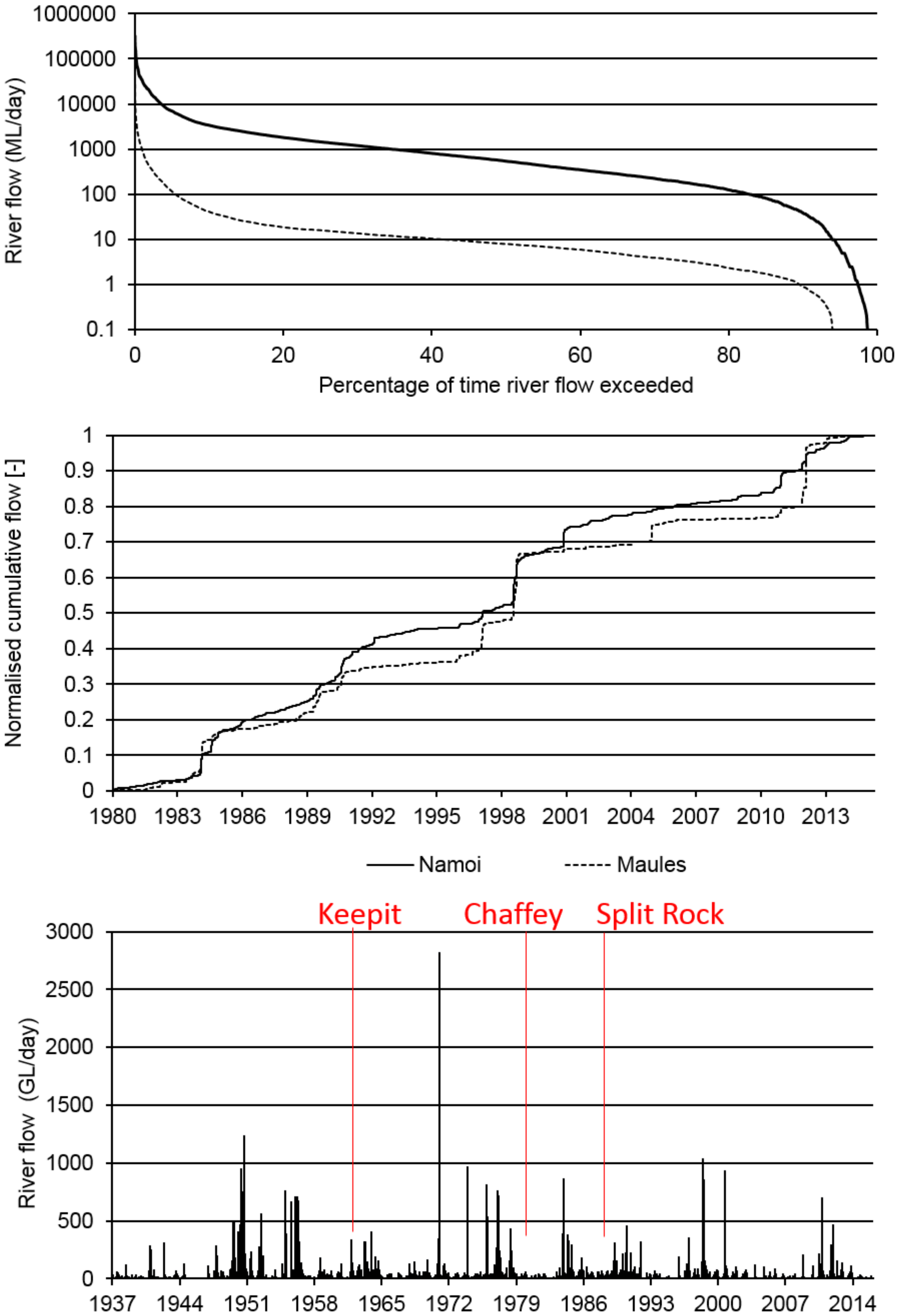


Figure 4.8 (a) Flow duration curve showing the difference of the flow and low-flow period of Namoi and Maules; (b) the normalised cumulative flow curve of the Namoi River and the Maules Creek; and (c) monthly river flow of the Namoi River showing the reduction of flood peaks after the completion of dams.

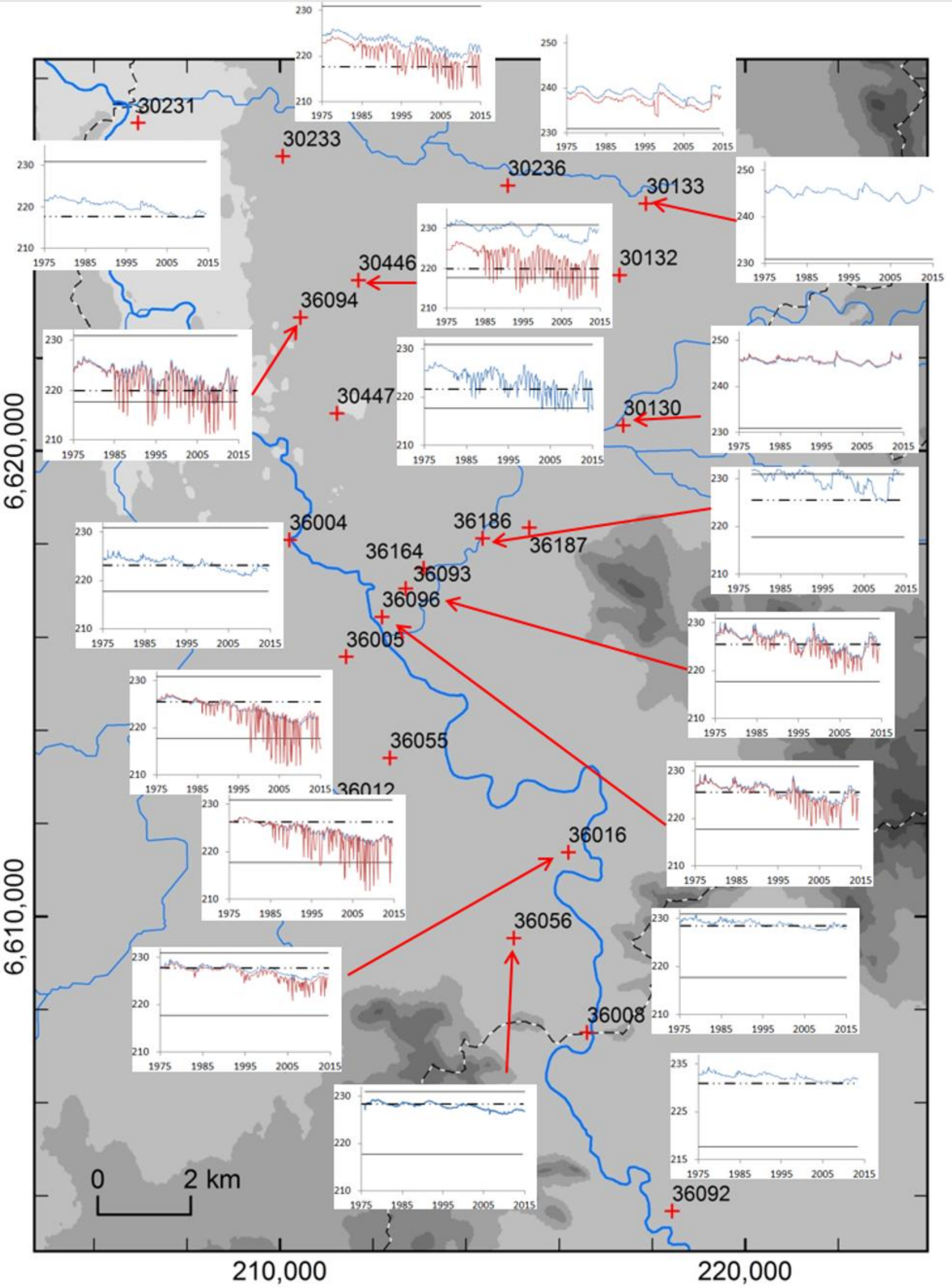


Figure 4.9 Groundwater hydrographs of selected boreholes in relation to their location. Data from 1975 to 2015 are shown in the inserts, with the same vertical scale. Hydrographs for the shallow aquifer (<30 m) are shown in blue and the deep aquifer (>30 m) in red.

rainfall in the floodplain was 66%, 66%, 58% and 62% of the long-term average. Wet periods occurred in 1984–85, 1996–98, 2005 and 2009–13, with the annual rainfall being 125%, 143%, 153% and 151% of the average, respectively. The most recent drought series commenced with the reduced rainfall in the early part of 2001 and peaked in late 2002, with 9 consecutive months of below average rainfall. Although the climatic drought was slightly relieved by extreme rainfalls in December 2004 and June 2005, the hydrologic drought condition continued through 2006 to 2007. The subsequent wetter period featured a series of months with over 100 mm rainfall in the summers between 2008 and 2013.

4.3.2 Stream discharge

Both the Namoi River and Maules Creek are characterised by infrequent floods and high streamflow events. Zero flow constitutes 1.1% for the Namoi and 5.7% for Maules. The discharge from Maules Creek is two orders of magnitude smaller compared with the Namoi due to its significantly smaller catchment area. From the flow duration curve in Figure 4.8 (a), flow events in excess of 5 GL/d occur approximately 90% of the time in the Namoi River. During dry years when little or no flow is recorded at Maules Creek, flow in the Namoi River is largely supplied by dam releases (Figure 4.8 [b]). Three major dams constructed in the headwaters of the Namoi River in 1960, 1979 and 1987 regulate the flow pattern of the Namoi River by maintaining the base flow and reducing the magnitude of flood peaks (Figure 4.8 [c]).

4.3.3 Groundwater levels

The groundwater level in the catchment generally follows the surface topography, which is higher in the east near the mountain compared to areas along the river. Potentiometric surfaces derived from the groundwater records show a general westward flow from the mountain front towards the Namoi River. Since the early 1980s, each dry period has led to a new low record in groundwater levels. During each wetting-up period, general increases in groundwater levels are observed across the catchment. Taking the most recent re-wetting in 2010–13 as an example, 57 out of the 61 monitoring boreholes in the catchment showed an increase of over 1.5 m from the lowest recorded level during the preceding drought. The average increase is 3.1 m with a standard deviation of 1.5 m. However, the increase varies spatially. Near the

Namoi River (Zone C), increases are less than 2.5 m, with the lowest increase (0.7 m) observed on the western river bank in the upstream catchment. In contrast, increases generally exceeded 2.5 m between the river and the mountains (Zone B). The highest increases were observed at the edge of the floodplain near the slope of the mountain (Zone A). A moderate increase was found near the perennial reach of Maules Creek, while there was a rise of up to 7 m in the ephemeral reaches of Lower Maules Creek. This gain in groundwater level gradually dissipated after the peak of the wet period. At the end of 2014, the average gain from the lowest levels of previous dry periods was 2.2 m, with a standard deviation of 1.1 m. Groundwater levels in Zone A retained their substantial gains. Lower in the Namoi Valley (Zone C), most boreholes lost more than half of the gain that has been lost since mid-2014.

4.3.4 Recovery ratios

Recovery ratios from major re-wettings are presented for 1984–85 (**Figure 4.10**), 1996–98 (Figure 4.11) and 2010–13 (Figure 4.12). These recovery periods were observed following the break of drought when the rate of precipitation exceeded evapotranspiration and provided opportunities for groundwater recharge. Zone A recovered to or above pre-dry period peaks with recovery ratios higher than 1. In contrast, levels along the Namoi River in Zone C did not receive enough recovery in the wet period. During these wet periods, events at 2.5 to 3 times greater than the mean monthly rainfall have been observed (Figure 4.6). A wet period with an above-average rainfall of 12 months or longer is required for the recovery to occur. In contrast, sporadic rainfalls within a prolonged dry period bring relief to the dry condition, such as the events in 1982, 1995 and 2005. However, the relative short duration is not enough to revert the general downward trend or bring catchment-wide recovery (Figure 4.13). Note that the slope of the CRD curve is generally used to indicate an upward or downward trend over a short period. Whether the values of the curve are positive or negative, is highly dependent on the antecedent conditions, and does not have any physical meanings; therefore, it should not be read as a direct surplus or deficit (Weber and Stewart, 2004).

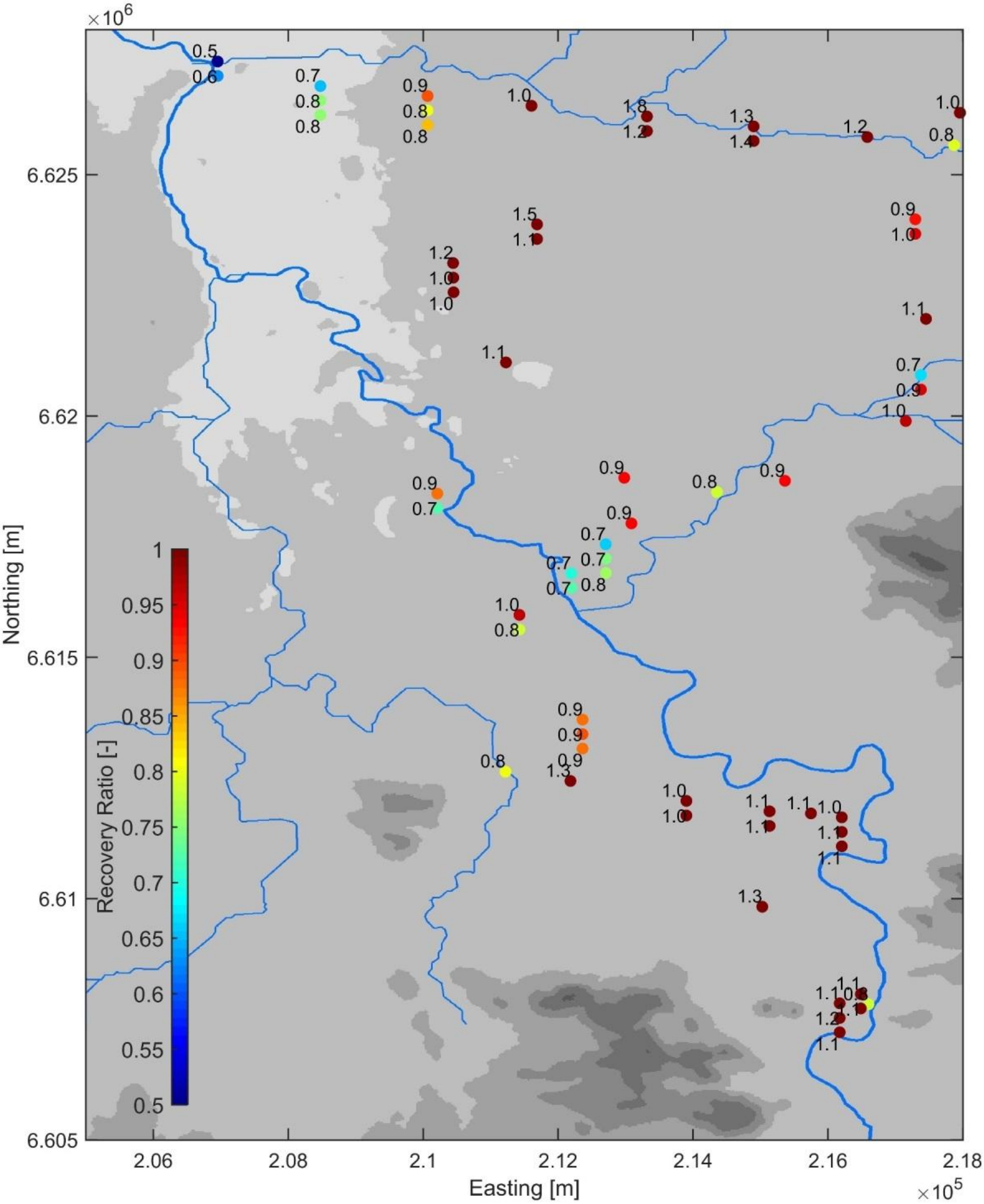


Figure 4.10 Recovery ratio for re-wetting between 1984–85

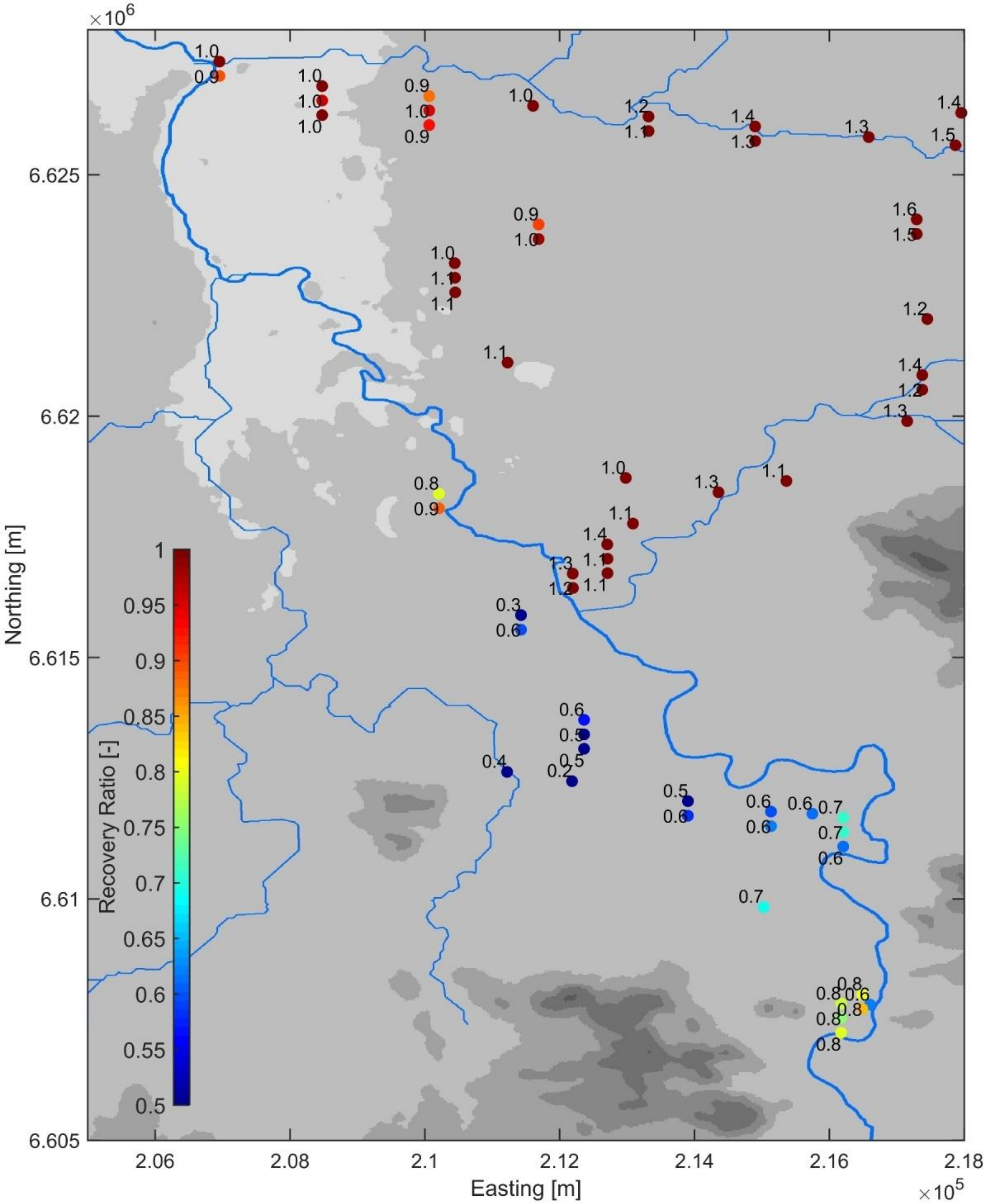


Figure 4.11: Recovery ratio for re-wetting in 1996–98.

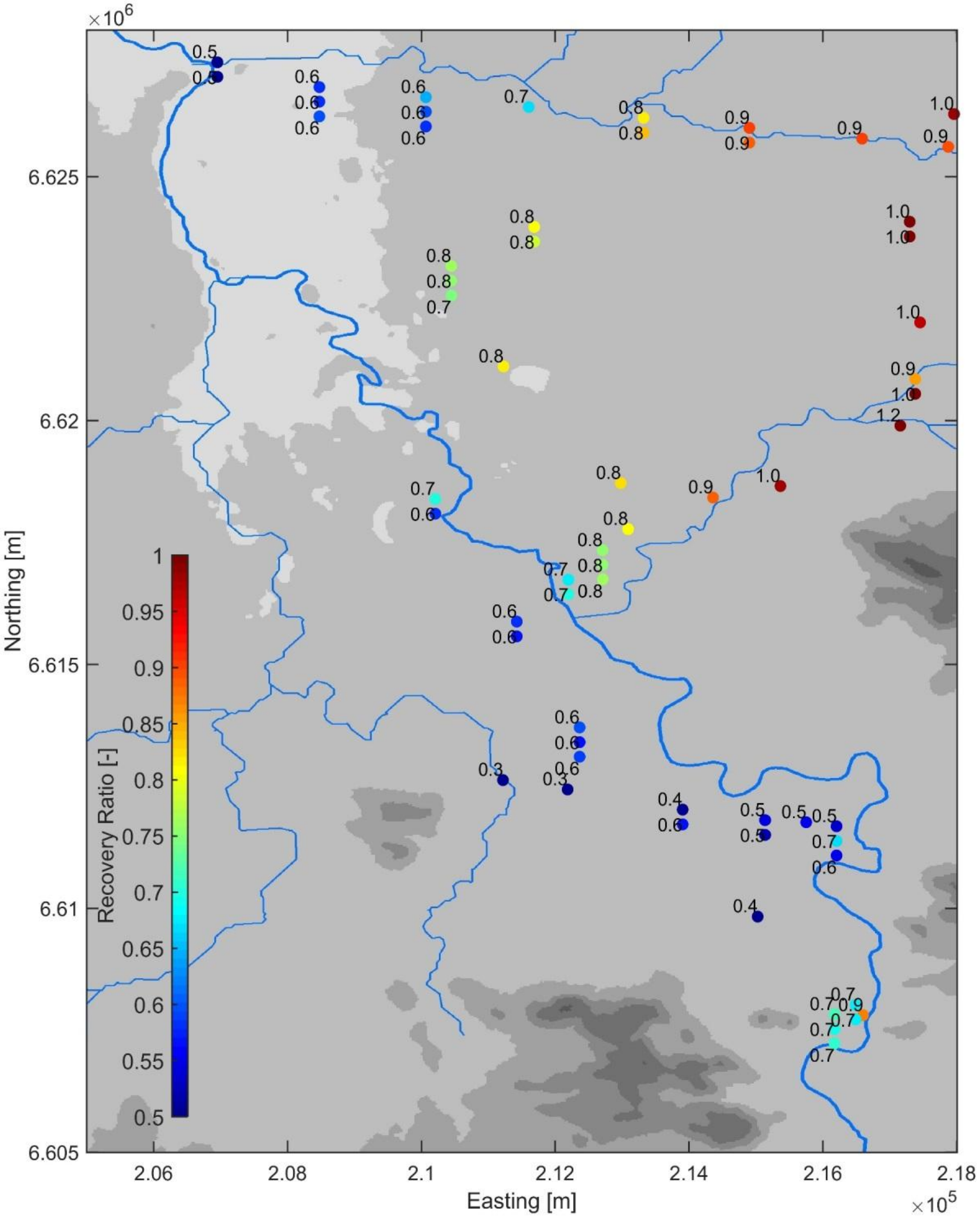


Figure 4.12 Recovery ratio for re-wetting in 2010–13. The pre-dry period peak after 1998 was used.

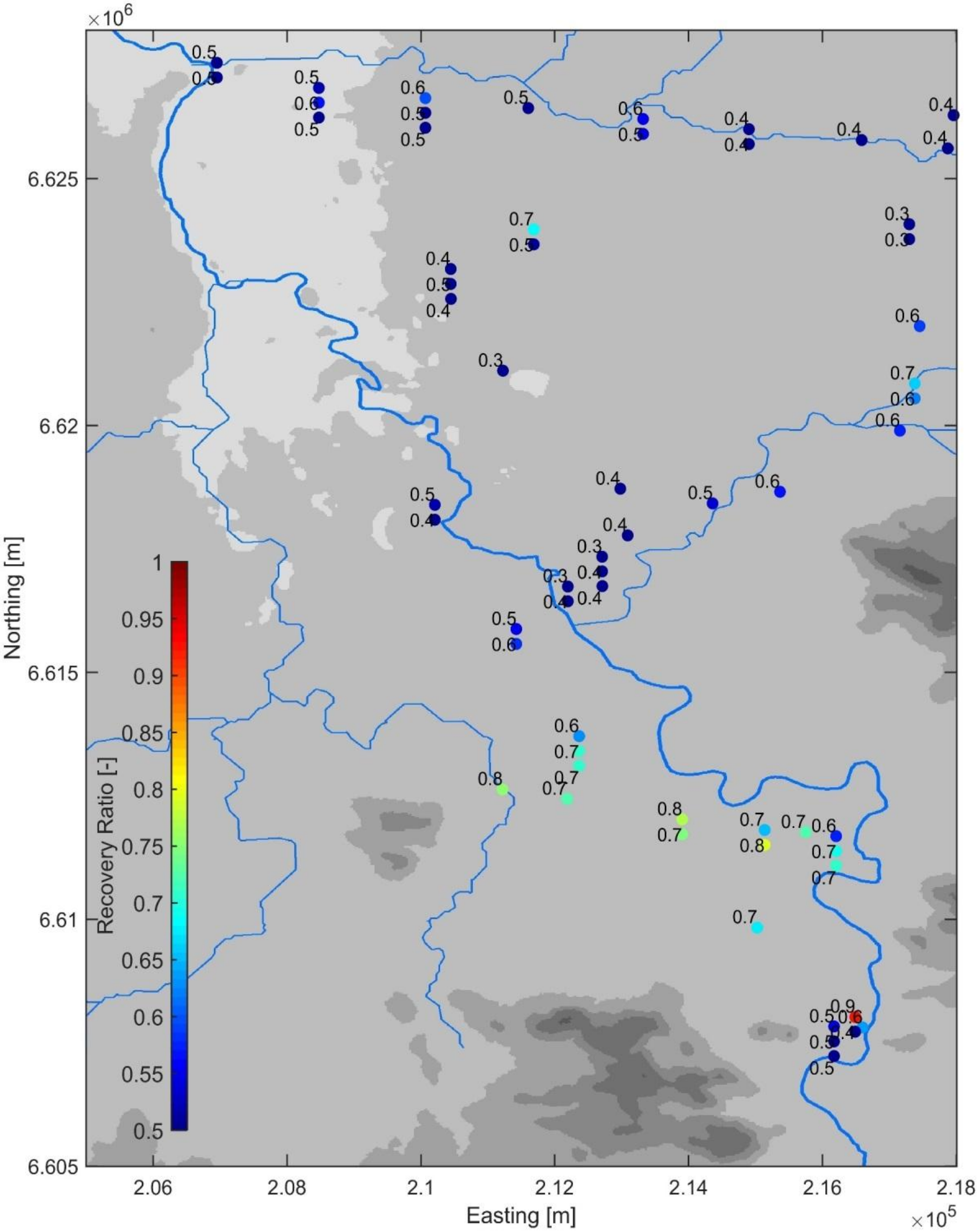


Figure 4.13 Recovery ratio for re-wetting in 2005.

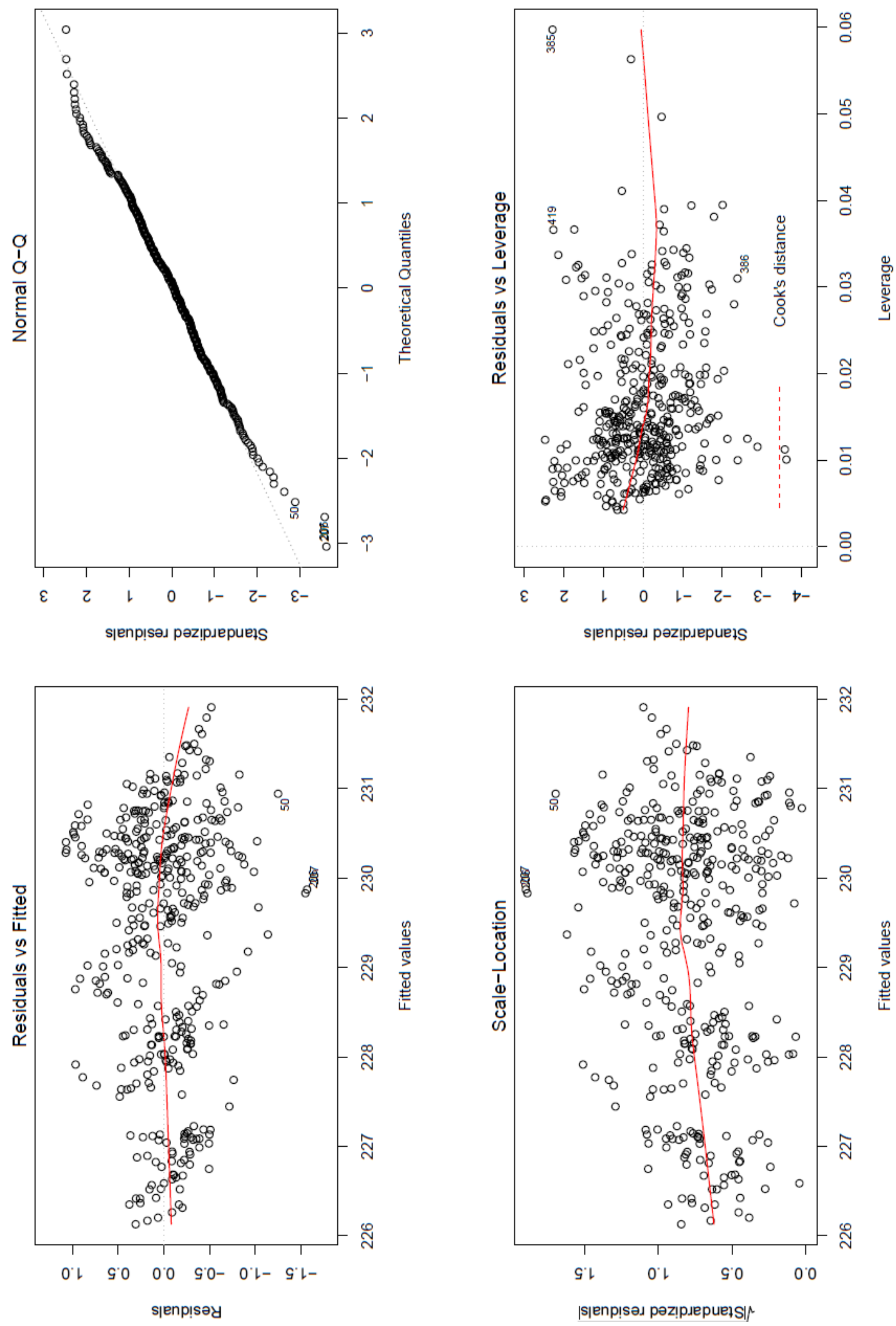


Figure 4.14 Linear regression model validation results for GW030446_1. Note: the corresponding r^2 is 0.9074.

4.3.5 Regression analysis

The optimised factor combinations of linear mixed models for describing each groundwater level time series are presented in Table 4.1. As indicated in the F-test, all optimised models achieved statistical significance ($p < 0.001$) in explaining changes in groundwater levels. An example of the statistical validation of the linear regression model is provided in Figure 4.14. The resultant r^2 ranged from 0.526 to 0.967 (Table 4.1).

For each optimised model, parameters that have statistically significant contributions ($p < 0.001$, where p is the probability in the t-test) are marked with ticks in Table 4.1. Although evapotranspiration was initially included in the analysis, t-test statistics showed no significant influence on any model ($p > 0.05$).

Most groundwater levels were shown to have statistically significant contributions by rainfall in the mountain range and the plain (Rain_MTL) and as surface water flow in Maules Creek (SW_Maules). Boreholes in Zone A are also affected by climate variation (represented by the Southern Oscillation Index [SOI]) and background long-term trend (represented by Year). In addition to long-term trends, Zone B boreholes are also characterised by seasonality (represented by Month). Surface water flow in the Namoi River (SW_Namoi) was generally influencing groundwater levels in Zone C boreholes only.

Table 4.1 Results of regression analysis with resultant r^2 of each model. Statistically significant parameters are ticked.

Borehole ID	R^2	Rain MTL	Rain TUR	SW Namoi	SW Maules	SOI	Month	Year	Zone
GW030129_1	0.7837	✓	✓		✓	✓	✓		A
GW030130_1	0.5259	✓	✓			✓		✓	A
GW030130_2	0.6697	✓			✓	✓	✓	✓	A
GW030131_1	0.6796		✓	✓	✓	✓		✓	A
GW030132_1	0.6861	✓	✓	✓	✓	✓			A
GW030132_2	0.71	✓	✓	✓	✓	✓			A

Borehole ID	R^2	Rain MTL	Rain TUR	SW Namoi	SW Maules	SOI	Month	Year	Zone
GW030133_1	0.7535	✓		✓	✓	✓			A
GW030134_1	0.7411		✓	✓	✓				A
GW030235_1	0.9007	✓		✓	✓	✓	✓	✓	A
GW030235_2	0.8786	✓			✓	✓	✓	✓	A
GW030236_1	0.8498	✓		✓	✓	✓			A
GW030236_2	0.8804	✓		✓	✓		✓	✓	A
GW030237_1	0.8401	✓			✓	✓	✓	✓	A
GW030232_1	0.9523	✓			✓	✓	✓	✓	B
GW030232_2	0.9268	✓			✓	✓	✓	✓	B
GW030232_3	0.9402	✓			✓		✓	✓	B
GW030233_1	0.9441	✓			✓			✓	B
GW030233_2	0.9329	✓			✓		✓	✓	B
GW030233_3	0.922	✓			✓		✓	✓	B
GW030234_1	0.9074	✓			✓			✓	B
GW030446_1	0.9103	✓			✓	✓	✓	✓	B
GW030446_2	0.9067	✓	✓		✓	✓	✓	✓	B
GW030447_1	0.8685	✓			✓	✓	✓	✓	B
GW036003_1	0.8934	✓	✓		✓	✓	✓	✓	B
GW036093_1	0.8228	✓			✓	✓	✓	✓	B
GW036093_2	0.8641	✓			✓	✓	✓	✓	B
GW036093_3	0.8547	✓			✓	✓	✓	✓	B
GW036094_1	0.868	✓	✓		✓	✓	✓	✓	B
GW036094_2	0.9035	✓	✓		✓	✓	✓	✓	B

Borehole ID	R^2	Rain MTL	Rain TUR	SW Namoi	SW Maules	SOI	Month	Year	Zone
GW036094_3	0.9092	✓	✓		✓	✓	✓	✓	B
GW036164_1	0.8941	✓	✓		✓	✓	✓	✓	B
GW036185_1	0.9494	✓		✓	✓	✓	✓	✓	B
GW036186_1	0.7493	✓			✓	✓	✓	✓	B
GW036187_1	0.7922	✓		✓	✓	✓	✓	✓	B
GW030231_1	0.9677	✓		✓	✓	✓	✓	✓	C
GW030231_2	0.9667	✓		✓	✓	✓	✓	✓	C
GW036004_1	0.9283	✓	✓	✓	✓	✓	✓	✓	C
GW036004_2	0.9506	✓		✓	✓	✓	✓	✓	C
GW036005_1	0.8997	✓			✓			✓	C
GW036005_2	0.9623	✓	✓	✓	✓	✓	✓	✓	C
GW036007_1	0.8987	✓		✓		✓		✓	C
GW036007_2	0.9132	✓		✓		✓		✓	C
GW036007_3	0.9038	✓		✓		✓		✓	C
GW036008_1	0.8894	✓		✓		✓		✓	C
GW036012_1	0.9558	✓			✓			✓	C
GW036013_1	0.9556	✓			✓			✓	C
GW036014_1	0.9547	✓	✓	✓	✓			✓	C
GW036014_2	0.9501	✓		✓	✓		✓	✓	C
GW036015_1	0.9444	✓		✓	✓			✓	C
GW036015_2	0.9379	✓		✓	✓	✓	✓	✓	C
GW036016_1	0.9551	✓	✓	✓	✓	✓		✓	C
GW036016_2	0.9426	✓		✓	✓	✓	✓	✓	C

Borehole ID	R^2	Rain MTL	Rain TUR	SW Namoi	SW Maules	SOI	Month	Year	Zone
GW036016_3	0.9456	✓		✓	✓	✓	✓	✓	C
GW036055_1	0.9557	✓	✓	✓	✓	✓	✓	✓	C
GW036055_2	0.9375	✓	✓		✓		✓	✓	C
GW036055_3	0.9547	✓		✓	✓	✓		✓	C
GW036056_1	0.8942	✓	✓	✓	✓			✓	C
GW036057_1	0.9067	✓	✓	✓		✓		✓	C
GW036057_2	0.9045	✓		✓		✓		✓	C
GW036096_1	0.848	✓		✓	✓	✓	✓	✓	C
GW036096_2	0.8788	✓			✓	✓	✓	✓	C

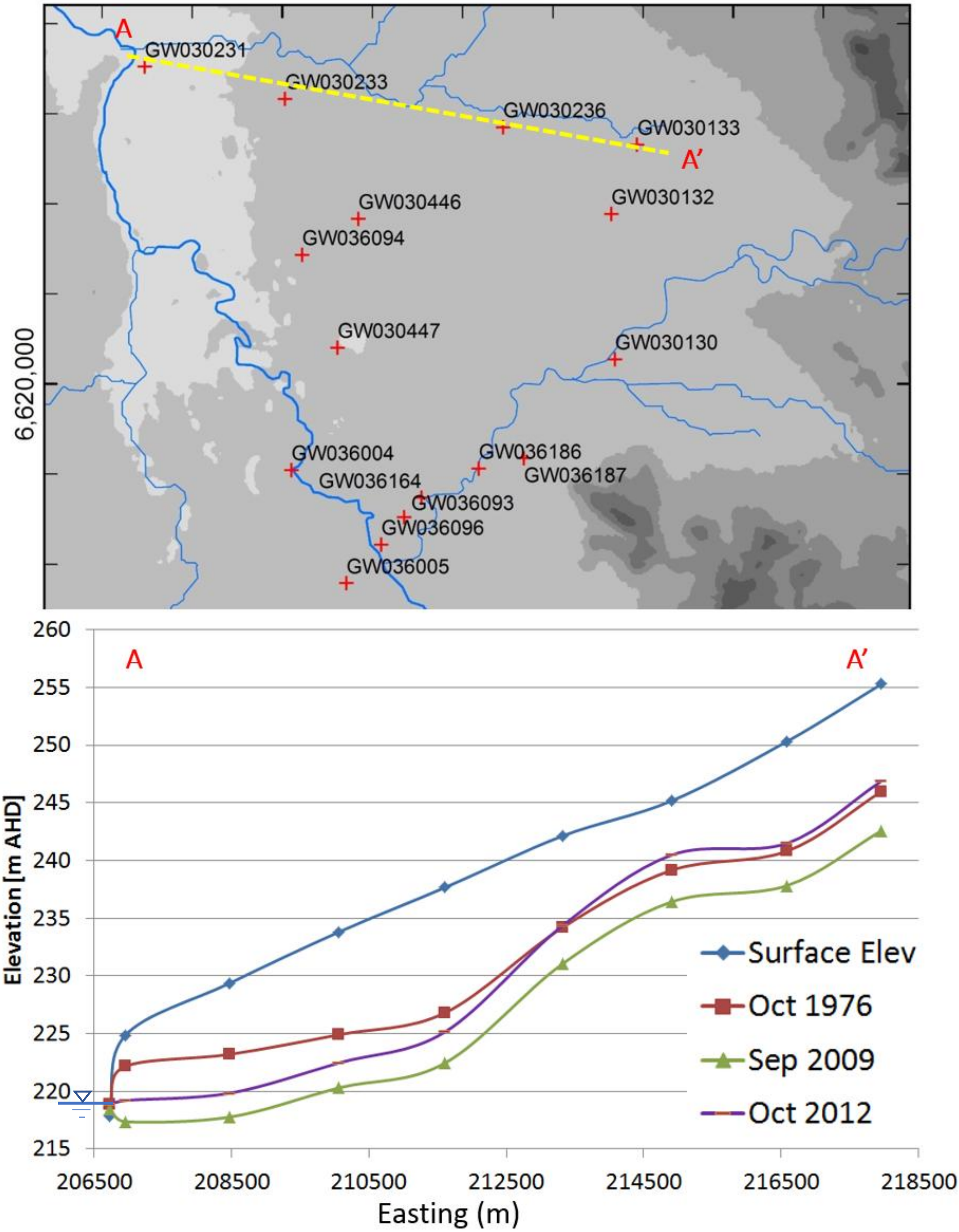


Figure 4.15 Profile of groundwater levels along the transect at the north of the catchment. Leftmost point is the water level in the Namoi River.

4.4 Discussion

4.4.1 Groundwater responses from stream recharge

Under natural conditions prior to the onset of large-scale pumping, the alluvial aquifer was dominated by the regional groundwater discharge from mountain slope infiltration and ephemeral stream leakage (Giambastiani et al., 2012). As shown by the pre-pumping groundwater level in Figure 4.15 (red line for October 1976 groundwater level), the hydraulic gradient towards the river drives the discharge into the Namoi River and contributes to the baseflow. The head difference between river and GW030231 (the borehole next to the river) reflects the lower river bed elevation compared with its surroundings. Such difference is also likely maintained by the low horizontal hydraulic conductivity near the river, which impedes the quick balancing of the groundwater level near the river.

With the onset of groundwater abstraction, the re-distribution of groundwater head under the alluvial plain altered the flow pattern of the groundwater. Evidence of pumping activities can be found in seasonal drawdowns in Figure 4.9 (Zone B), which are approximately situated along the Namoi River paleochannel. As the total abstraction rate exceeds the amount of regional discharge from the mountain area, the deficit is shown as a drawdown of the groundwater table in the alluvial plain (McCallum et al., 2013). Initially, pumping mainly captured regional groundwater discharge. For parts of the Namoi River, base flow contribution from groundwater discharge reduced or even ceased. During dry periods when more abstraction occurred, such as in 2009 (as shown by the green line in **Figure 4.15**), the lowered groundwater level also caused the loss of river flow.

The recovery of groundwater heads observed over the past few decades indicate that groundwater recharge during a wet period serves as a source of replenishment for groundwater reservoirs. Out of the 61 groundwater hydrographs analysed in the regression analysis, rainfall in the mountains provided statistically significant contributions to all but two locations. On the other hand, only 20 hydrographs were found to have significant contributions from rainfall on the plain. The spatial distribution of the recovery ratio for the three recorded wet periods also showed a decreasing trend from the Nandewar Ranges towards the Namoi River (from east to west). It is understood that rainfall on the mountain is higher (Figure 4.2) due to the orographic effect and, thus, rainfall occurring on the mountain is more likely to form recharge. Infiltration in Zone A through the coarse-grained materials in the colluvium area at the interface between the mountain ranges and the alluvial plain provides the source of the regional groundwater flow. This accords with the event analysis in Chapter 3 and the work by Giambastiani et al. (2012).

The amount of precipitation during these wet periods should be large enough to overcome the soil deficit in the upper catchment to create flows in the ephemeral and intermittent stream channels. Moreover, consecutive wet years are required to allow the equilibration of the groundwater level following individual storm events. In this process, ephemeral stream reaches provide the recharge pathways necessary to allow water to infiltrate the streambed during and after high rainfall periods and spread water laterally through the dissipation of the groundwater mound that has developed beneath the streams (Cuthbert et al., 2016). The recovery ratio of 2005 (Figure 4.13) provided a counterexample of how a short, wet period with higher rainfall cannot provide enough recovery at the catchment scale.

Contributions from streamflow of Maules Creek were shown to be a significant source of recharge. During flow events, infiltration through riverbeds and inundated areas replenishes groundwater if sediments are sufficiently permeable and causes a quick increase of the groundwater level near river channels that propagate laterally into the aquifer (McCallum et al., 2014, King et al., 2014). Under the intermittent reaches of Maules Creek, unsaturated zones can quickly become saturated and allow recharge to occur. The phenomenon has been described by Shanafield et al. (2012). Results of the regression analysis showed that streamflow in Maules Creek (SW_Maules) provided significant contributions to 54 out of 61 boreholes, including all but one borehole in Zone A and Zone B. The prevalence of the SW_Maules signal also in boreholes physically separated from Maules Creek (e.g., as to the west of the Namoi River ver) indicates that the SW_Maules signal represents a more general ephemeral stream recharge process happening in other streams.

The influence of the Namoi River is limited to areas near the river. The majority of Zone C boreholes (22 out of 27) showed contributions from streamflow from the Namoi River. Yet, very few Zone B boreholes (2 out of 21) showed contributions from the

Namoi River. Such findings suggest the influence of the Namoi River flow is limited to the lowest part of the catchment, approximately 2 km to the east of the Namoi River. Boreholes to the west of the Namoi River are also included in this zone of influence. Given that the topography to the west is relatively flat, this area does not have any mountain front input as in the eastern part. Therefore, flow in the Namoi River is the dominating hydrological driver for this part of the catchment. In contrast to groundwater levels affected by mountain front recharge from the Nandewar Ranges, aquifers to the west of the Namoi River (Zone C) received limited recovery of groundwater levels during the recent wet period (Figure 4.11 and 4.12). Observations indicate the groundwater levels continue to follow a downward trend over the last few decades, despite the presence of smaller peaks (relative to Zone A and B) in response to river floods. The background trend term in the regression analysis has statistically significant contributions to all boreholes, except a few in the upper catchment part of Zone A. Although flooding in the Namoi River provided temporal and localised increases in groundwater levels, such recharge either dissipates after the passing of the flood peak in two to four months or is not detected in the low-resolution record.

On a decadal scale, the recharge from the Namoi River streamflow events had little effect on the trend of the groundwater level. This is shown by lower recovery ratio values (< 1) observed during all three recorded wet periods. Based on the records between 1980 and 2015, the period of rise and fall of the groundwater levels is consistent with the wet/dry climatic periods and is therefore unlikely to be due to long lag effects. These findings indicate flood or stream recharge may only be temporary along rivers. The hydrograph trend along the Namoi River for the wetting-up period between 2010–2013 accords with the event analysis for the connected stream reaches detailed in Chapter 3 at EC and BV. At locations with connected conditions, groundwater and surface water are in a dynamic equilibrium. Although during dry periods the groundwater level falls below the base of the river and forms a losing system, the fact that the groundwater and river levels are connected allows only limited storage space in the shallow aquifer. Meanwhile, areas of major groundwater abstraction are further away and at a higher elevation compared to the river. Although high river stages during floods may create hydraulic gradients in favour of aquifer recharge, the infiltration is temporary and, therefore, is mainly bank storage, most of which returns to the river after the peak recedes (Vázquez-Suñé et al., 2007).

4.4.2 Implications for potential managed aquifer recharge using event discharge

It was acknowledged that, during the dry period, the level of groundwater extraction in the Namoi Catchment exceeded total long-term groundwater recharge and, therefore, was regarded as not being sustainable (CSIRO, 2007). Given the scale of established agricultural development in the region, demands on groundwater resources will remain for the foreseeable future. Groundwater level observations along the Namoi River during this moderate to wet study period indicated that abstractions in the alluvial plain have mainly captured regional groundwater discharge. During extended dry periods, when groundwater levels dropped below the river stage, leakage from the Namoi River provided a small but continuous recharge to the aquifer and stabilised the groundwater level near the river relative to areas further away from the river (**Figure 4.15**). Such reversal from naturally gaining stream conditions to pumping induced losing conditions to cause additional stream leakage and further reduce streamflow and change the water balance for a catchment (Braaten and Gates, 2003, McCallum et al., 2013).

The increasing agricultural water demand in semi-arid areas requires more water storage for growing season irrigation (Bond et al., 2008). To meet this demand, traditional water management approaches of storing water in dams have been extensively applied in Australia and worldwide throughout the 20th Century (Larsen et al., 2014, Peirson and Laut, 2016). The forecasted increase in water demand is unlikely to be met with new dams since optimal reservoir locations are already exploited in most regions (Scanlon et al., 2016). Moreover, the excessive water loss through evaporation reduces the available water to meet the demand and reduces the economic viability of storing water in surface reservoirs (Hassan et al., 2015).

It has been observed in the Lower Namoi that the construction of flood mitigation structures and irrigation networks increased the recharge during floods (Ross et al., 1991). Structures across rivers, such as weirs, also increase groundwater recharge near these structures due to streamflow retention. At Mollee Weir, which is approximately 15 km downstream of Narrabri on the Namoi River, a mean infiltration rate of 0.29 m/day from the river to the underlying aquifer has been identified (Kelly et al., 2009, Lamontagne et al., 2011). The unexpected beneficial side effect of the weir is the increased sub-surface water storage of infiltration through the river bed when the river is flowing. The principle is similar to Managed Aquifer Recharge (MAR), which is to engineer the storage of water in aquifers for subsequent use (Dillon et al., 2009). In rural areas, MAR may be useful for replenishing water in aquifers subject to over-exploitation or to improve water quality in saline aquifers (Dillon, 2005). Storing water in aquifers requires less surface area of the operational footprint (vs large dams) and is free from evaporative losses observed in surface water storage. Examples of MAR have been reported in Australia and overseas to lower cost and environmental impact compared with surface storage (Khan et al., 2008, Choi and Lee, 2012, Raju et al., 2013, Singh, 2014).

Since groundwater is generally of good quality at the Maules Creek Catchment (Andersen and Acworth, 2009), adopting MAR of flood waters would have to exploit locations with unsaturated zone storage, as there are limited options for MAR operations displacing poor groundwater with better quality MAR water.. Results from this study suggests that infiltration along the disconnected reaches of ephemeral and intermittent streams could be providing significant groundwater recharge compared with the connected reaches of Maules Creek at Elfin Crossing and along the Namoi River. Consequently, it is possible that an additional amount of streamflow in the ephemeral streams during high-flow events can be captured and infiltrated with recharge weirs. Recharge weirs, also known as percolation tanks, are designed to increase the residence time of streamflow in the channel during streamflow events and to allow water to infiltrate through the bed, therefore increasing groundwater storage in the unconfined aquifers. The water captured adds to regional groundwater flow and can reduce the drawdown in the lower part of the catchment. An example of such application in the Minderoo Station in Western Australia is shown in **Figure 4.16**. The following practical elements should be considered when selecting potential locations for recharge weirs (Dillon et al., 2009):

1. *An adequate source of water for recharge.* Ephemeral creeks of 3rd order or greater that receive storm runoff from mountain ranges can be considered. This to ensure any potential MAR site has a suitably large upstream catchment size. For example, the 4th order reach at MC described in Chapter 3 carries flow for over two months following large storm events, which can potentially be utilised for MAR projects.
2. *A suitable aquifer to store the water.* Recharge weirs should be built in reaches over a significant extent of unsaturated materials. As shown for the detailed

study sites in Chapter 3, unsaturated aquifers near MC would work better than in Maules Creek near EC. In addition, the lower part of Maules Creek that intersects the paleochannel from the east may also provide a recharge pathway to the underlying aquifer.

3. *A good connectivity between the stream and the aquifer.* Suitable groundwater monitoring observation is required to identify aquifers that quickly respond to streamflow events. Other than MC, sites with quick groundwater response also included Lower Maules Creek near governmental monitoring borehole GW36186 and Old Bibbla Creek near GW30236.

The practical aspects of implementing MAR in the Maules Creek Catchment or in similar semi-arid areas requires further investigation. Consideration of the impact to downstream water users may also need to be assessed. For example, water captured and held or used in excess of what can be saved from evaporating at the surface (streams or dams) would be otherwise available for downstream users. A detail analysis of the economics or costs-benefits of implementing the MAR technology is outside the scope of this study. More detailed discussion in the literature includes Dillon et al. (2009), Khan et al. (2008) and Maliva (2014). In addition, the following concerns need to be addressed in future studies:

i. For all unsaturated aquifer systems, there is an upper limit to how much water can be captured during a particular event, or during a wet period comprising a set of events. It has been shown at MC (Chapter 3) that, once the stream becomes connected, the recharge reaches an upper limit. For such conditions, only a very limited amount of additional streamflow may be captured. The limit should be quantified to determine the efficiency of such a project.

ii. Not all streamflow events onto Maules Creek and its tributaries lead to significant discharge all the way into the Namoi River. For smaller events, all the flow ends up infiltrating into the alluvial aquifer through the streambed. Capturing more of the streamflow in the upper catchment will therefore potentially reduce the available volume of water for sustaining surface flow in the lower reaches of Maules Creek. Thus, with MAR structures in place, the number of downstream events and their flow duration may decrease. The potential ecological consequences, then, need to be carefully considered.

iii. The scale of such a MAR project suggested above is relatively small scale compared to the potential flow in the Namoi River. As shown in section 4.3.2, during streamflow events, available water in the Namoi River is 1–2 orders of magnitude larger than in Maules Creek. However, utilising event flows in the Namoi River and in the Maules Creek Catchment by installing river weir structures may not be as effective due to the limited thickness of the unsaturated zone below the river. Consequently, implementation of MAR would require more complex engineering structures, which may include a series of pipeworks or channels, and pumping stations and infiltration facilities to transport and store water in zones with considerable unsaturated aquifer thickness. Suitable areas could be the zones with groundwater level drawdown due to abstraction in the paleochannel to the east of the Namoi River. The capital and maintenance costs would be much higher than for simple weirs across the river.

iv. Implementing MAR may lead to water user equity issues between upstream and downstream water users. From a holistic perspective of the water balance for the entire Murry Darling Basin, the total amount of water is not increased through MAR, other than the reduction of the potential evaporation loss from the alternative option of storing equal amounts of water in surface reservoirs. Therefore, drought-proofing of the Maules Creek Catchment may come at the expense of the lower Namoi and the Darling system. Further research would be required to optimise the goals of protecting priority ecosystems and irrigation farming throughout the Namoi and for downstream portions of the Murry Darling Basin.

Figure 4.16 Recharge weir constructed across the Ashburton River at Minderoo, Western Australia (Koenigluck, 2015).

5 Concluding remarks

The primary motivation for undertaking this research was to identify the processes involved in the groundwater recharge through stream channels in a semi-arid area during a climatic transition from dry to wet conditions. Due to significant groundwater monitoring infrastructure, the Maules Creek Catchment in the Namoi Valley, NSW, Australia, is a good example in studying the change in hydrological processes due to climate variations. The Nandewar Range to the east of the catchment provides significant storm runoff, which drains into a series of intermittent creeks. While the hydrogeology and water balance of the catchment has been studied previously, considerable uncertainties are associated with determining the dominant form of recharge and its quantification. Furthermore, a better understanding of the surface water-groundwater interaction as a function of climate variability will be essential for sustainable water management in the future with predicted climate change.

The research was divided into two parts. First, a detailed analysis of key monitoring sites with high frequency data (15- to 30-minutes observations) along a hydrologic and geomorphic gradient from the mountain front to the floodplains, with ephemeral and intermittent stream reaches in the upper part and perennial reaches in the lower parts of the catchment was performed. This part focused on the analysis of groundwater responses to individual streamflow events. Second, a catchment scale analysis of monitoring records from 35 bores since the 1980s was conducted. This part served to identify long-term trends in groundwater levels over multiple climatic transitions from dry to wet.

It was possible to demonstrate that all the sites studied in the Maules Creek Catchment are influenced by high streamflow events based on responses in groundwater levels. However, it was shown that the surface water-groundwater interactions are more complex than is generally assumed. Although groundwater levels do respond to high streamflow, whether the aquifer can receive recharge will depend on the type of stream reach and the antecedent conditions. Along an ephemeral or intermittent stream reach, while a recharge pathway exists through the streambed material, the available space for accepting groundwater recharge depends on the groundwater level prior to the streamflow event. At the perennial stream reaches where surface water is in balance with its surrounding groundwater table, it is possible to observe an increase in groundwater level following a short duration, high stream stage. However, the water gained in the event is mostly bank storage, which stays in the riparian aquifer for a duration of only weeks to months before re-entering the stream as the stream stage recedes.

At the catchment scale, following the transition from the dry to wet period, multiple medium-to-large rainfall events (130–260 mm per month) in the headwater area caused run-off in the mountains and streamflow in the intermittent stream channels. Along the intermittent reaches of Maules Creek, where groundwater abstraction had created an area of lowered groundwater levels, transmission loss during flow events between 2010 and 2013 provided significant recharge and the water table rose to near or above pre-irrigation development (~1980) levels. Away from the intermittent streams, however, at the end of the recovery period, groundwater levels were still 0.5–2 m lower than pre-abstraction levels. Moreover, the section of the Namoi River in the study was historically naturally gaining, which has been weakened since the onset of pumping. The wet period merely re-balanced the river and aquifer in the vicinity of the river without fully re-establishing the pre-irrigation development gaining conditions. The results provide evidence that widespread groundwater level drawdown that is created during a prolonged drought is unlikely to be readily recovered by diffuse recharge. This is particularly observed in the lower catchment where there is a high water demand for flood irrigation. The analysis suggests that it may require years for regional groundwater flow from the upper catchment recharge zones to recover the drawdown.

The study highlights that streams can dynamically behave as ephemeral, intermittent or perennial depending on the climatic condition and antecedent groundwater levels. When assessing the impact of dry and wet periods on the amount of groundwater that may be used sustainably and still support ecosystem health, it is critical that both conceptual and quantitative water balance assessments correctly consider the switching of stream reaches depending on prevailing climatic conditions (wet or dry periods).

From a water resource management perspective, this study explores some aspects of the feasibility of managed aquifer recharge (MAR) for floodwaters along stream channels. In semi-arid areas similar to the Maules Creek Catchment where natural recharge may be slow, there is a potential for carefully designed MAR to enhance the groundwater resource and to reduce the periods of stress experienced by groundwater-dependent ecosystems. However, before the economics of managed aquifer recharge can be examined, there is need to refine our understanding of potential hydrological, ecological and engineering challenges in the implantation of the MAR of floodwater in highly dynamic stream environments.

6 Reference

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