Government of **Western Australia**
Department of **Water**

Drainage for salinity control at Pithara

Looking after all our water needs

Salinity and land use impacts series

Report no. SLUI 46
November 2010

Drainage for salinity control at Pithara

by

NM Cox

Department of Water

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Disclaimer

Cover photograph: The downstream drain gauging station at the Pithara site
Photographer: Nick Cox

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An Appendix CD is in the back cover of the printed version. For online readers this CD is available on request from the Salinity and Water Resource Recovery Branch of the Department of Water.

Summary

A single groundwater drainage scheme slightly reduced but was unable to totally prevent the risk of dryland salinity in a Wheatbelt trial funded by the Engineering Evaluation Initiative. While the drainage removed saline groundwater it did not sufficiently lower the watertable beneath the land targeted for salinity recovery.

Deep open drains have been constructed in many catchments in Western Australia and are increasingly seen as useful for draining groundwater with the aim of reducing land salinisation. As drain effectiveness remains unclear, the factors that contribute to the success or otherwise of these drainage schemes were not well understood and knowledge was not being transferred from one scheme to another.

The Pithara drainage project assessed the practicality of lowering the watertable beneath a saline valley floor for the recovery of agricultural land for dryland cereal cropping. In 2004, 18 kilometres of 2.5 m deep groundwater drain were dug to manage a growing salinity problem in the valley floor of a small Wheatbelt catchment east of Pithara. The mainly single drains wound their way along the valley floors and discharged directly into a naturally saline watercourse.

The drains contributed to small reductions in the watertable height that lead to some improvement in land condition but not enough to allow for dryland cereal cropping. To allow for cropping the watertable needed to fall an additional half to one metre below its original height.

Drain discharge water quality varied markedly with discharge rates. Highly saline and acidic discharges were associated with low flow conditions while higher flows both diluted salts and neutralised acidity. Of the total water drained during 2004–06, 85% was from groundwater with most of the surface water inflows originating from one severe storm event.

Watertable responses have been used to assess the efficacy of most drainage schemes in the Wheatbelt. At Pithara, watertable measurements alone were found to provide inadequate information on how groundwater responded to the drains. The watertable fluctuations alongside were the result of differing recharge, evaporation, lateral flow and drainage. It was impossible to directly measure the effects of the drain alone on lowering the watertable.

Only because drain discharges were accurately measured in conjunction with groundwater responses could relationships between the drain and its surrounding aquifer be more clearly understood. The volumes of drain discharges suggested the potential for far greater watertable reductions compared to those measured, even though drainage efficiency was limited by aquifer permeability.

The conceptual drain and aquifer water balance used in this report suggested significant limitations with the Pithara drains in that a single channel was unlikely to drain groundwater from its catchment faster than it was replenished. The hidden and unrealistic expectation was that the drains could de-water the regional-scale aquifer with only local-scale drainage efficiency. In light of this, it did not matter how efficient the drain was at removing groundwater, the surrounding watertable levels would not fall significantly.

To protect land from salinity the watertable must be lowered enough to reduce the capillary rise and capillary discharge of groundwater that saturates the plant root zone and transports salt to the soil surface. To date, the indicator of drainage success has been the horizontal extent of the influence of single drains on the surrounding watertable. For salinity control, what is really important and often overlooked is that watertable reduction is the precursor to success.

Modelling suggested that the upstream 9 km of the Pithara drain had capacity to lower and control watertables beneath 170 ha of saline land. To achieve this however, requires isolating the 170 ha groundwater catchment of the drain from the surrounding catchment. As unachievable as this sounds it is exactly the rationale and methodology used to design agricultural drainage worldwide.

Watertable control beneath irrigated agricultural land is conventionally achieved with drains dug parallel to each other. At construction, the distance between the drains is designed to maintain the groundwater level at or below a nominated depth while balancing drainage efficiency against recharge and the soil porosity. As the drains are parallel to each other, their groundwater catchment areas and recharge are readily calculated, enabling the catchments to be appropriately sized to match the efficiency of the drains.

Although not irrigated land, this same drain design approach and expected watertable responses are considered applicable to the Pithara project site. Based on the climatic variables and aquifer characteristics, modelling suggests that parallel drains with 150–250 metres spacing could lower and control the watertable enough to recover once-saline land for dryland cereal cropping.

Most importantly, the project highlighted that drainage design is ineffective if undertaken in isolation of setting clear performance objectives and assessing the achievement of these within the context of the whole-of-catchment groundwater system. A thorough assessment of drain performance is based on consideration of all of the potentially affected variables of the water balance. Taking this approach to drain design will likely lead to major changes to the current ‘single channel’ approach to Wheatbelt drainage.

1 Introduction

The water balance in the Wheatbelt in Western Australia was changed by clearing deep-rooted perennial vegetation for agriculture. Groundwater recharge has increased in some areas to about 10% of annual rainfall. This is causing groundwater levels to rise, resulting in damaging levels of salt accumulation in plant root zones. Dryland salinity mainly affects the poorly drained broad valley floors of the Wheatbelt. This saline area is expected to expand to over three million hectares by 2015 (State Salinity Council 2000).

In the 1970s saline watertables began to approach the soil surface and landholders began to advocate and construct drains to increase discharge and manage rising watertables and waterlogging (Coles et al. 1999). Constructed at the farm-scale, drains were dug 1–3 m into the soil to intercept groundwater, and thus became commonly known as ‘deep drains’.

Forty years on the total length of drains and banks in the Wheatbelt exceeds 11 000 km (ABS 2003). Yet, there remained a general lack of knowledge and information on the effectiveness and impacts of drains and drainage schemes. Most of the evidence of saline land reclaimed with drainage was anecdotal. Discharges into existing natural waterways were fuelling environmental concerns about the quantities and quality of the waters involved.

In 2004, the Government of Western Australia initiated the Engineering Evaluation Initiative (EEI) through the now Department of Water. The main scope was to focus on increasing understanding of the appropriate use of engineering options to manage dryland salinity for economic, social and environmental benefit (Dogramaci & Degens 2004). As part of the EEI, the Department of Water conducted a cooperative trial with landholders in the north-eastern Wheatbelt to investigate a groundwater drainage scheme.

The project was to design, construct, and evaluate a farm-scale, deep, open drainage scheme that was typical of other Wheatbelt schemes. The groundwater conditions and drain discharge volume, salinity and chemistry were measured before and for two and a half years after construction.

The objective of the monitoring program was to address as much as possible all aspects of the scheme’s water balances as a prerequisite to understanding the effects of the drain on the surrounding watertables and groundwater systems. The results have been confirmed and extended by spreadsheet modelling and using the steady-state drainage equation.

1.1 Objectives

The main aim of the Pithara EEI project was to evaluate the effectiveness of a deep open drain on lowering the watertable, and the resultant discharges from the drains. The secondary aim was to investigate the difference in discharge responses between leveed and non-leveed drains. The measurements from the project were also used to support partnership projects to investigate the recovery of saline land in response to drainage (Bell et al. 2009), and the downstream impacts of drain discharge into the environment (Strehlow et al. 2006).

To meet the main objectives a monitoring program was developed to identify and measure as far as possible the aspects of the water balances that affected the drainage scheme.

The specific study components were:

- Define the aquifer characteristics at the site that affect drainage.
- Calculate the water balances at project scale.
- Better understand the drivers of watertable responses to the drain.
- Investigate the effectiveness of the drain in lowering the watertable.

1.2 Supporting information

Supporting information is provided with this report on an Appendix CD. Despite frequent reference to this CD in the text, access to it is not considered essential in terms of the results discussed within this report. The Appendix CD contains supplementary information that includes:

- All of the measured data used in this report
- Detailed plans of the project site
- Drilling logs
- Supporting spreadsheets for the data analysis and interpretation

1.3 Errors and accuracy

The high number of significant figures provided for much of the data and analysis within this report should not be misconstrued as indicating an associated high degree of accuracy. In most cases, numerical results are presented with a high number of significant figures to enable the reader to follow the evolution of, and interrelationships between results.

It should be noted that, even in the absence of gross errors, individual measurements contain inherent inaccuracies associated with instrument error and field techniques. Suggested levels of error for the variables measured for the Pithara project and used in this report are:

- Groundwater level measurements: +- 0.05 m
- Drain discharge (by stage): +- 20%
- In-situ salinity: +- 5%
- In-situ pH: +- 0.5 units
- Continuous salinity: +- 10%
- Rainfall: +- 5%

The systems approach used in this report is expected to have reduced the potential for the development of erroneous conclusions from the measured results. The systems approach is based on the use of data trends and interrelationships, rather than focusing the results from individual parameters.

This report has deliberately steered away from using complex hydrological models in this initial assessment of the Pithara drainage project. This has been done in order to identify and gain consensus of the main variables that affected the performance of the drain and watertable responses at Pithara. This more simplistic approach is also more likely to be replaceable which may not be the case for complex modelling, where input variables can mask underpinning data, assumptions and relationships.

2 Pithara drainage site

The Pithara drainage project is approximately 200 km north-east of Perth on the property of Kingsley and Paula Roach within the Shire of Dalwallinu. The property, Petrador Farms, is 25 km ESE of Dalwallinu in the Western Australian Wheatbelt (Fig. 1). The agricultural enterprises conducted on the property are dryland cropping and the grazing of livestock typical of the Wheatbelt. The selection of this project site was supported by the Dalwallinu Land Conservation District Committee.

The site was selected to represent a typical north-eastern Wheatbelt property with low relief, poorly defined natural drainage and broad valley floors with mainly clay subsoils. Most of the catchment surrounding the project area extends beyond the approximate 3600 ha Petrador Farms property that abuts the drains. The total catchment drained is approximately 13 200 ha, of which 1037 ha was slight to moderately and 278 ha severely affected by dryland salinity.

The project focused on constructing and monitoring 18 700 m of drain that crosses the property in a north-westerly direction. Approximately half way along the drain it passes under the Pithara East Road at map coordinates (MGA 50) 490 800 mE, 6 638 700 mN.

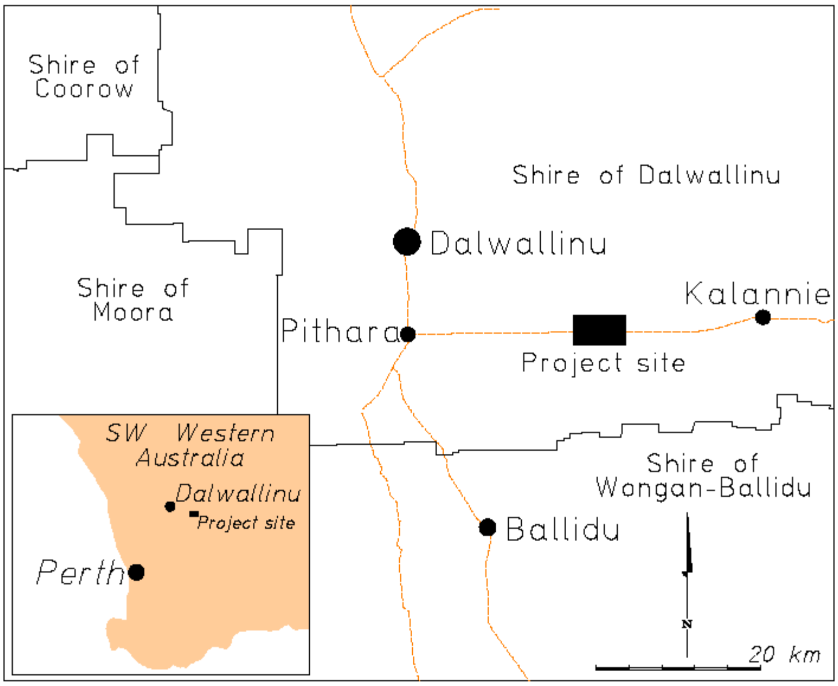

Figure 1 Project site location within the shire of Dalwallinu

2.1 Climate

The climate is temperate with cool wet winters and hot dry summers. Bureau of Meteorology records for the period 1957–2008 show the Dalwallinu town site (25 km NE of the project site) has average minimum and maximum temperatures of 17.4–33.6 °C during the warmer months December–March and 7.3–18.8 °C for the cooler months June–September. An average of 119.2 days a year had maximum temperatures higher than 30 °C with 11.1 days over 40 °C.

The average annual rainfall for Dalwallinu (based on 96 years of record 1912–2008) was 356.5 mm and ranged from 119.9 mm (in 1914) to 672.4 mm (in 1999). The annual rainfall for the last 37 years (Fig. 2) averaged 353.1 mm, ranging from 198.6 mm (2002) to 672.4 mm (1999). The unusually high 1999 rainfall is well above the next highest recorded (512.1 mm in 1984) and is largely attributed to the March rainfall of over 200 mm.

Approximately 60% of the rain falls during May–August. The weather systems that bring winter rains are low pressure systems that originate in the south-west and travel eastward over the south of the continent. These rains affect large geographic areas and are generally of low intensity and long duration. They are more frequent and regular than the summer rains which are mainly from the tropical north-west and can be more extreme and localised.

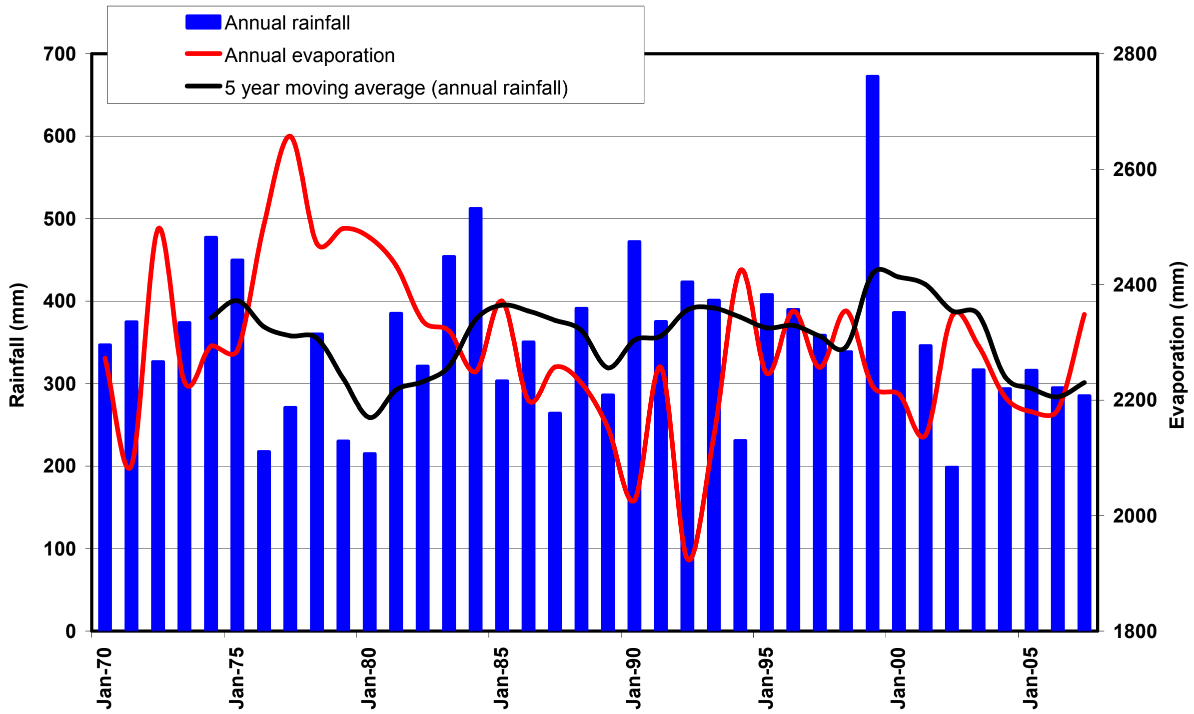

Figure 2 Annual rainfall and evaporation at Dalwallinu (1970–2007)

The rain gauge (Station 508041) was operational at the drainage site from July 2004 to January 2007. A total of 690.9 mm of rainfall was measured during the 925 days of record, including 249.9 mm and 296.7 mm respectively for 2005 and 2006. The most significant rainfall (139.2 mm) during January and February 2006 was from several consecutive daily events in excess of 20 mm (Fig. 3).

Despite the high rainfall at the start of 2006, it was still thought that the 2005 and 2006 rainfalls measured at the site were below average, based on a comparison with previous and the 2005–06 rainfall at Dalwallinu. Rainfall at the site was 66.1 mm less than the below-average rainfalls for Dalwallinu during 2005, and 2 mm more than in Dalwallinu in 2006.

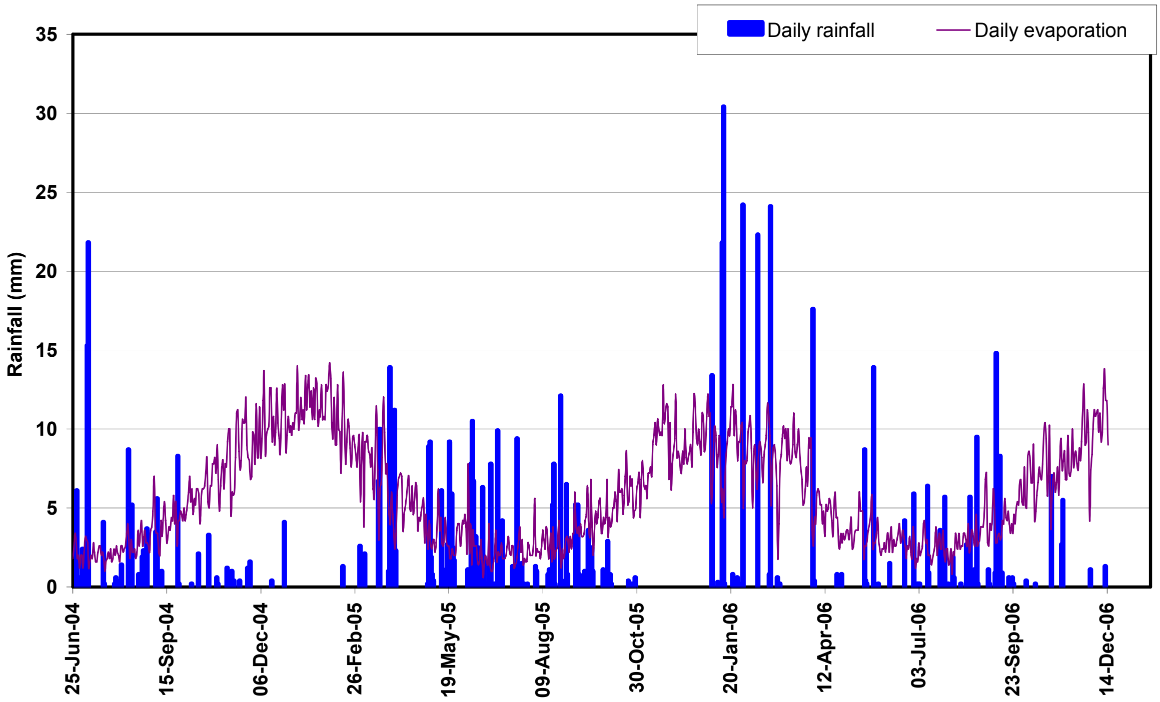

Figure 3 Daily rainfall and Silo evaporation at the drainage site (Station 508041)

In the July 2004 to January 2007 period, total evaporation at Dalwallinu was 5486.6 mm (SILO 2006). Silo evaporation is data interpolated from evaporation measurement sites surrounding a particular site to produce a composite estimation. Peak summer evaporation during the life of the project was 14.6 mm/d during December 2006 (Fig. 3). Evaporation rates reached slightly higher than 20 mm/d between 1970 and 2004. High evaporation rates are not uncommon in the Wheatbelt and are usually exacerbated by hot dry spring and summer easterly winds from the inland.

Watertable levels and drain discharge appear most responsive to a combination of significant daily rainfall events and monthly rainfall trends (Cox & Tetlow in press). Accumulated monthly residual rainfall (AMRR) is used to best reflect monthly rainfall trends within a given period. AMRR is the progressive accumulation of rainfall for each month less the average monthly rainfall for the period of analysis.

During the normal summer winter rainfall seasons AMRR declines from around October and rises again from about April (Fig. 4). Large increases in AMRR outside this seasonal trend were mostly in response to significant (>30 mm/d) summer and early winter rainfall, as in January 2006 (Figs 3 & 4).

An analysis of AMRR for Dalwallinu from 1970–2007 showed a general decline since 2000 (Appendix CD 2.1). The average monthly rainfall 2000–07 was 25.4 mm, 5 mm less than the 1970–2000 average, reflective of the declining AMRR. The 1970–2007 rainfall for Dalwallinu tends to indicate that the post-2000 declining rainfall trend is at least partially related to a reduction in rainfall events in excess of 30 mm/d. Annual rainfall prior to 2000 included on average at least one rainfall event in excess of 30 mm/d, with many of these in excess of 40 mm/d. From 2000 to 2007 there were only four rainfall events in excess of 30 mm/d.

Because watertable levels appear to correlate with both AMRR and individual larger rainfall events, the absence of larger events is expected to produce falling AMRR trend which could be reflected in falling watertable levels. Groundwater levels across the northern Wheatbelt may be in decline since 2000, in response to the lower rainfall (Speed & Kendle 2008). The inference is that natural watertable decline could be attributed to a drain response. While it is acknowledged that the dry seasons may also have some influence on the drained watertable, the response time frames to drainage compared to natural decline should produce sufficiently differentiated results.

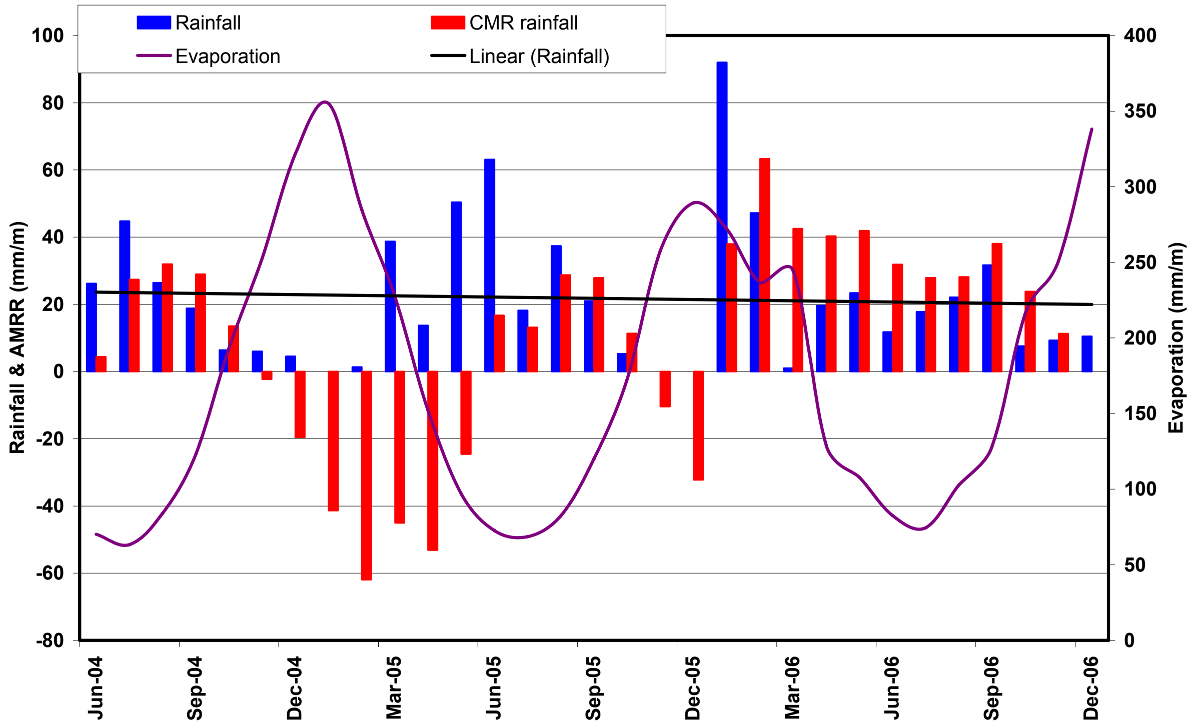

Figure 4 Monthly and AMRR at the drain site and SILO evaporation from Dalwallinu

2.2 Land use

Europeans settled at Dalwallinu and Pithara to develop wheat and sheep farming and planted the first crops in 1907. The current agricultural land use is predominantly dryland cropping of cereals and legumes from autumn through to spring. Livestock (mostly sheep) graze crop residues, introduced pastures and perennial shrubs, with numbers subject to seasonal influences and livestock prices. Only 12% of the native vegetation remains in the 595 418 ha agricultural zone of the Dalwallinu Shire (Shepherd et al. 2002).

Petrador Farms retains 239.1 ha (6.6%) of native vegetation on the landholdings that surround the project site. Since clearing, mostly native halophytes have colonised the valley floors and associated areas as they became progressively salt-affected. The halophytes include mostly the volunteer shrubs bluebush (*Maireana* spp.) and saltbush (*Atriplex* spp.) (Fig. 5).

The salt-affected and halophyte-covered land provides limited economic return beyond that derived from the opportunistic grazing of livestock. The Pithara drain was constructed on land almost entirely salt-affected and colonised by halophytes.

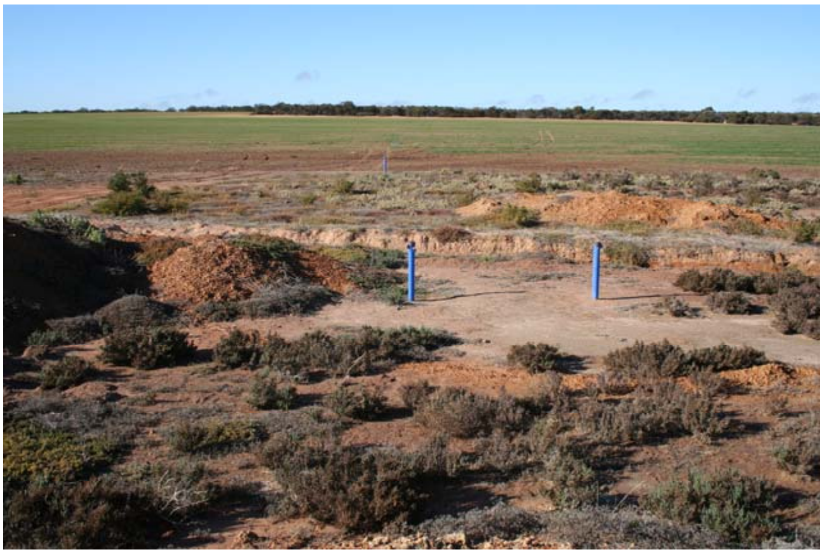

Figure 5 Salt-tolerant vegetation on the valley floor and footslope salinity on transect 2

2.3 Topography and natural drainage

Petrador Farms is just within the northern upper watershed boundary of the North Mortlock River. The Mortlock drains the upper north of the Swan Avon drainage basin. Channel formation is poorly integrated with the river consisting mostly of salt lake chains that drain west and then south from the property. Due to the semi-arid climate streamflow within the upper Mortlock and its tributaries is ephemeral, being active only after high intensity or prolonged rainfall events. Streamflows are usually highly saline to brine, with salinities of 14 000 to more than 35 000 mg/L (Mayer et al. 2005).

Petrador Farms is characterised by low relief adjacent to the broad central valley floor that is the focus of this drainage project. Valley flanks typically consist of long slopes at 1–3% rising to 60 m above the valley floors. The valley flanks abruptly give way to a flat (often less than 0.1%) valley floor 200–2000 m wide, with several small drainage depressions or clay pans (Fig. 9 in Section 2.6). The valley floor landscape enters the south-east of the property at 320 m AHD (Appendix CD 3.1), crosses Pithara East Road and exits the north-western boundary at 296 m AHD.

Drainage discharge from the scheme was measured by two gauging stations (Section 3.2). The catchment of the upstream gauging station (615044) is 6165 ha, encompassing most of the property south of Pithara East Road (Fig. 10 in Section 3.1). It rises from 297 to 400 m AHD at about 8 km SE of the gauging station. The upstream catchment is a subcatchment of that of the downstream gauging station (615045). The total catchment of the downstream gauging station is 13 200 ha. The drain outlet at the western boundary of the property is a further 1 km downstream from the gauging station (Fig. 10).

Gradients along the valley and drains are very low. The average gradient along the 6 km watercourse between the gauging stations is 1:1500. Runoff from the property landscape is ephemeral, requiring significant and/or sustained rainfall to overcome the losses inherent with the flat valley floor landscape. The salinities of the runoff from the valley floors are consistent with the salinities elsewhere within the upper North Mortlock River.

2.4 Soils

McArthur (1991) describes the soil landscapes as dissected laterites and lateritic sandplain. Most of the valley floor soils are grey-brown deep loamy duplex (Schoknecht 1997), with deep sandy and deep sandy duplex soils formed on the valley flanks. Isolated areas of saline wet clay soils are associated with ephemeral groundwater seepage close to the centre and at some points along the footslopes of the valley floor.

The typical soil profile near much of the drain is red–brown or yellow–brown loamy and silty sand over a mottled medium to heavy clay subsoil (Fig. 6; Bell et al. 2009). Much of the structure of the upper profile is massive and the subsoil, pedal. Red loamy sand textured mottles increased in number and size with depth to occupy about 30% of the profile at 2 m. In the more inundated and salt-affected parts of the valley floor mottles became cemented or the cemented material combined to form red-brown hardpans at about 2.0 m below ground surface.

Silcrete was excavated along about 1 km of drain upstream from the transect 2 bores (Fig. 10). This material occurred from about 1 m below the ground surface and was sufficiently hard at one location to resist excavation necessitating realignment of a section of the drain. Neither the hardpans nor silcrete appeared to form intact layers or structures that would prevent the movement of groundwater. Clayey material above the hardpans often contained coarse quartz fragments with the clay itself having an open (porous) fabric while in-situ. Other than in the presence of the silcrete the drain was excavated with relative ease.

Figure 6 Typical soil profile of the Pithara drain showing loamy sand over mottled clay subsoil

Calcium carbonate nodules, up to 50 mm in diameter and occupying up to 2% of the soil profile were found at 0.2–1.0 m depth along some south-eastern sections of drain. The soils and groundwaters in test pits at these sites were neutral to alkaline (pH 7–9) compared to those at the other sites that were acidic (<6.5).

The loamy and silty sand topsoils and upper soil profiles were more predisposed to erosion than the medium to heavy clay subsoils. In the more saline environments the combination of soil erosion caused by dispersion and slaking, and windblown material quickly silted up some of the channels. Channel silting was most noticeable where the drain was open to surface water runoff (Fig. 7).

Figure 7 Saline and erosion-prone soils and channel silting at Pithara

2.5 Geology and hydrogeology

The drainage catchment is within the Archaean Yilgarn Craton and comprises mainly gneiss, granitoid and migmatite basement rock (Carter & Lipple 1982). During the Tertiary Period, deep weathering produced an extensive lateritic profile subsequently partially or completely eroded from most of the catchment. Undifferentiated Tertiary and Quaternary alluvial deposits occupy most of the valleys.

Regolith varies in depth and composition, usually comprising one or more of:

- slope deposits, including colluvium and sheet-wash
- sandplain, mainly aeolian, including some residual deposits
- residual or relict material, including ferruginous, siliceous and calcareous duricrust
- exposed rock, saprolite or saprock.

Basement rock outcrops close to and along some of the catchment divides. The largest of these outcrops is Petrudor Rocks which covers an area of about 10 ha, 8 km ESE of the upstream gauging station (Fig. 8).

Figure 8 Runoff pooling below Petrudor Rocks

Twenty-seven investigation and groundwater monitoring bores drilled in the valley floor south of Pithara East Road (Section 3.2) included a 4 km transect of four deep bores to basement rock at 14.5–29 m. From the drill logs (Dogramaci 2004) the regolith is broadly described as having three layers:

- The bottom layer, approximately 2 m thick, developed above the basement by the formation of angular medium to coarse grains from the fragmental disintegration of basement rocks, rich in quartz and feldspar.

- The middle layer is the clay-rich layer embedded with minor sand and quartz grains that is the weathered profile. The clay is generally white tending upwards to pink and extends to about 5 m below the ground surface. Green clay at the interface of the two deepest layers in bore PT001d (at 12–13 m) and PT012d (at 19–27 m) may indicate where this layer is semi-permeable to upward movement of groundwater from the deep layer to the surficial sediments that cover the majority of the valley flats. Yet, no significant difference in water level between the deep and surficial aquifers was measured.

- The top layer is the 1–2 m thick surficial sediments, of medium to coarse-grained quartz sands and clays overlie the in-situ weathered profile. Surficial deposits derived from alluvial, colluvial and aeolian origins appear to have a higher hydraulic conductivity than the clay-rich layer. Indurated silcrete was encountered in some bores between 2 and 5 m and might indicate a historically high watertable.

Hydraulic conductivity (K_{sat}) was measured directly in four shallow bores using the slug withdraw method (Freeze & Cherry 1979). These were the second bores from the drain in each of the four bore transects (Fig. 10). The hydraulic conductivities for three of the bores were 0.015 to 0.056 m/d with an average of 0.036 m/d and a standard deviation of 0.02 m/d; the fourth was 0.308 m/d in bore PT008.

Hydraulic conductivity (K_{sat}) was also estimated prior to drain construction using a number of 2.0–2.7 m deep test pits along the drain alignment. Hydraulic conductivity estimated from groundwater inflow rates into seven of the pits, averaged 0.018 m/d (standard deviation of 0.017 m/d). This figure was used to support the drain design and construction approvals processes.

Groundwaters are found in all of the three layers of the regolith, with little indication of groundwater confinement within or between them (Section 4.1). In combination, the aquifer lithology of the three layers is dominated by weathered granite (Dogramaci 2004). Although not measured, the permeability of the weathered granite aquifer is expected to vary widely. Hydraulic conductivities of 0.01–3.99 m/d have been measured in similar aquifers elsewhere in the Wheatbelt (George & Frantom 1988).

Previous drilling on the property confirms that the weathered granite extends from beneath the valley floor, under the valley flanks, sometimes thinning at the margins of the basement rock outcrops. Hillside aquifers developed within the weathered granites and sometimes within the overlying sandplain are thought to discharge, contributing to groundwater recharge beneath the valley floors and seepages along the footslopes (Fig. 5).

2.6 Watertables and salinity

Nulsen (1998), and Speed and Kendle (2008) differ in their assessments of recent watertable changes beneath the Wheatbelt, possibly reflecting the timing of these in relation to rainfall and recharge trends. Watertables may currently be declining on the valley flanks due to recent dry seasons, but not in the valley floors. Ghauri (2004) notes that deep groundwaters continue to rise in some lower slope and valley positions indicating groundwater heads have been maintained despite recent dry conditions. Groundwater rises of >0.1 m/yr are still occurring beneath lower slopes and valleys where recharge areas are dominated by sandplain slopes.

The continuing spread of dryland salinity across the valley floors shows that rain-fed recharge and head-driven groundwater rise are still in effect across Petrador Farms. In 2004, the watertable beneath about 960 ha was mapped as being within or rising to within 1 m of the land surface (Fig. 9). Groundwater levels across the more salt-affected areas of the valley floor were 0.5–1 m below ground level, increasing in depth with both increasing elevation along the valley floor and distance away from its centre line (Appendix A).

Salinity caused by seepage from the sandplain valley flanks is clearly evident along many footslopes (Fig. 5) and is most noticeable at the confluence of minor hillside drainage depressions with the valley floor. At these points, the abundance of seepage water can prolong the growing seasons for crops and pastures, as evident by the late season green patches within the otherwise matured (brown) crops (Fig. 9).

The depth to groundwater progressively increases from near ground level at the footslopes (Appendix A.3) to potentially tens of metres below ground level on the valley flanks and hilltops. Existing bores and the recent drilling show the depth to groundwater to be influenced by the thickness of the weathered profile and comparative thickness of the hillside aquifer developed within the profile. Although not measured, in some instances a perched aquifer may develop in the sandplain surficial sediments overlying the weathered profile (Section 2.5).

Groundwater levels under the valley flanks remain consistently above the elevation of those beneath the valley floor and maintain hydraulic gradients towards the footslopes. In this report, hillside aquifers are viewed as a source of recharge and constant head to the valley floor aquifer. Hillside aquifer discharge to the valley floor aquifer contributes to the largely static groundwater levels beneath the valley floor, as noted by Ghauri (2004).

When the watertable rises to close to the land surface beneath the valley floor, upward groundwater rise is countered by groundwater evaporative loss from the land surface. The evaporation of saline groundwater leaves salts to accumulate on the land and within the topsoil. The most severely salt-affected areas on the property overlying both high watertables and the most saline groundwaters can become white salt scalds completely denuded of vegetative cover (Fig. 9).

Valley floor groundwater salinities on and surrounding the property are generally saline and can exceed 36 000 mg/L (seawater). In contrast, very fresh hillside sandplain seeps at some footslopes contain less than <275 mg/L and are used as livestock water supplies (Nulsen 1998) or may be drawn upon by crops and pastures.

There is a strong inverse relationship between depth to watertable and the surface expressions of salinity on the property. Areas severely affected by salinity such as those low on the valley floor are associated with saline watertables less half a metre from the surface. This follows a general trend of increasing groundwater salinity associated with decreasing elevation (Section 4). Salinity generally increased in moving from both the footslopes towards the centre, and from the upper towards the lower parts of the valley floor.

Watertable salinities measured on the property ranged from around 3000 mg/L beneath footslopes to 74 000 mg/L beneath the severely salt-affected land surrounding the downstream gauging station. Salinities close to the upstream drain alignment ranged from 21 000 to 33 000 mg/L while those at 400 m alongside were 5000–15 000 mg/L.

Although less predictable in its distribution, groundwater acidity also tended to increase downgradient along the valley floor. The pH levels were consistently less than 3.0 pH units in bores surrounding the lower valley floor on the western part of the property.

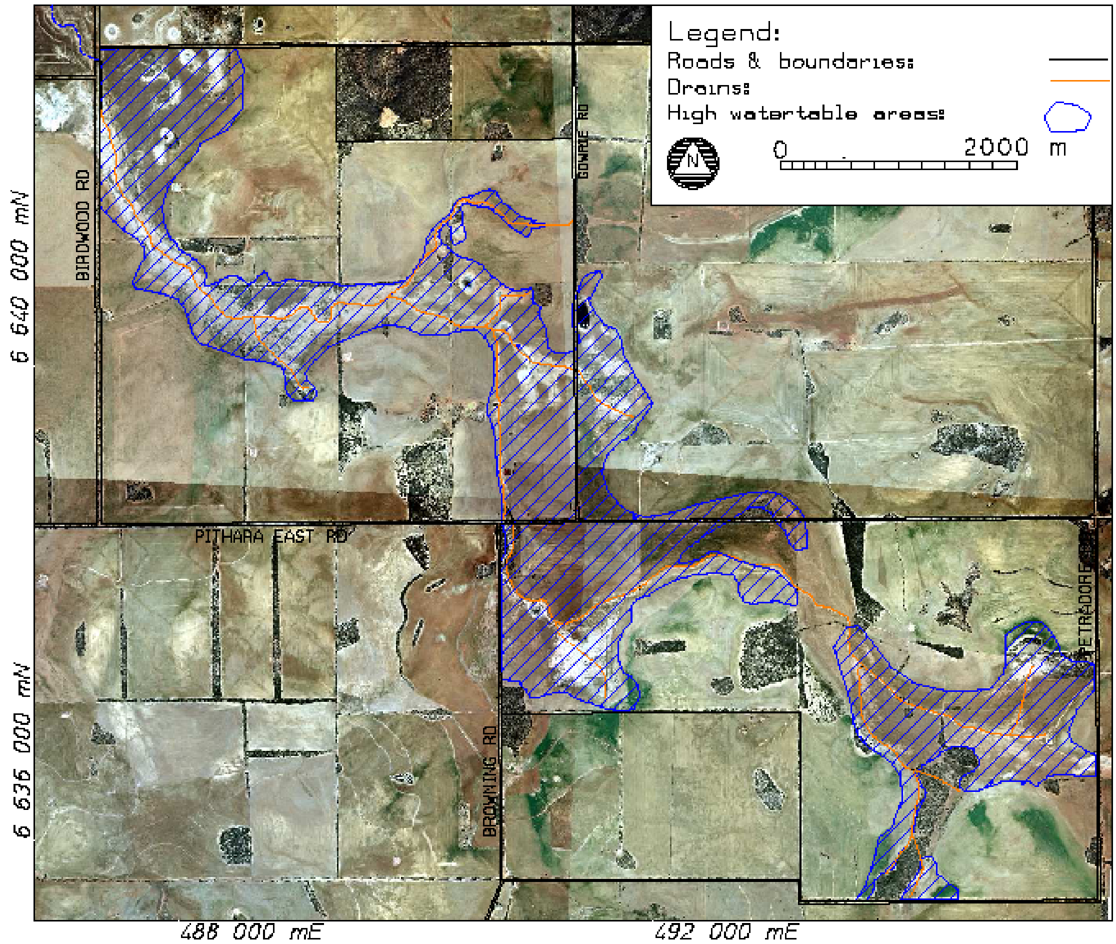

Figure 9 Areas on the property with saline watertables rising to within 1 m of the land surface

3 Pithara project design and methods

3.1 Drain design and construction

By completion on 8 July 2004 a total of 18 675 m of drains were constructed above the downstream gauging station, comprising a main drain with seven smaller tributary drains (Fig. 10). Some of the drains were constructed before the EEI project including 2100 m of 2 m deep open groundwater drain and 2610 m of 1.2 m deep, 4 m wide excavated natural watercourse.

The construction of 13 960 m of 2.5 m deep open and leveed drains for the EEI projects took approximately 45 days in June and July 2004. A 51-tonne bucket excavator (Fig. 11) worked upstream along the surveyed (new) drain alignments (Fig. 10). The excavator bucket had a trapezoidal profile to allow it to cut the 0.5:1 batter slopes.

Excavation commenced from the upper end of the already excavated watercourse at a depth of 1.2 m, increasing in depth over 580 m to 2.5 m. The 2.5 m deep groundwater drains had a cross-sectional areas of 6.12 m^2. For drains on the north of the Pithara East Road spoil was placed to form discontinuous levee banks to allow surface water to flow into the channel at selected points. The downstream 2460 m of the 2.5 m deep main channel extended to just north of Pithara East Road. Part way along this section, a 1470 m tributary drain branched off in a south-easterly direction to drain saline land on the east side of Gowrie Road (Fig. 10).

The main channel extended 6225 m upstream from the Pithara East Road. Although excavated to the same specifications as above, the spoil was placed to form continuous levees along both sides of the drain to exclude surface runoff (Fig. 6). The 3810 m of tributary drains that joined the main channel were also completely enclosed within levee banks.

The newly constructed main drain was positioned as close as possible to the lowest alignment along the valley floor, and along most of its length, on the northern and eastern sides of a surface water drain. The surface water drain was approximately 0.3 m deep, and increased in width from 4 m upstream to 10 m downstream. Culvert crossings were provided over the groundwater drain for vehicles and livestock. For the leveed upstream drain these also allowed surface water to flow over the drain at ground level. All culverts are 600 mm in diameter, including the 40 m lengths under the Pithara East and Gowrie Roads.

As part of the design and approvals process, the Hoodghoudt steady-state equation (Ritzema 1994) was used to estimate the discharge volume and watertable drawdown for from the new drains. The input values included the drain dimensions and estimated recharge of 10% of average annual rainfall (35.6 mm/yr). Average hydraulic conductivity (K_{sat}) was 0.018 m/d (Section 2.5) combined with an estimated 15 m aquifer thickness beneath the drain (Section 2.4), and 0.75 m pre-drain watertable depth.

The total steady-state groundwater discharge into the new drains was estimated at 268 kL/d or about 19 kL/d/km. The calculated watertable zone of influence was about 110 m across the drain.

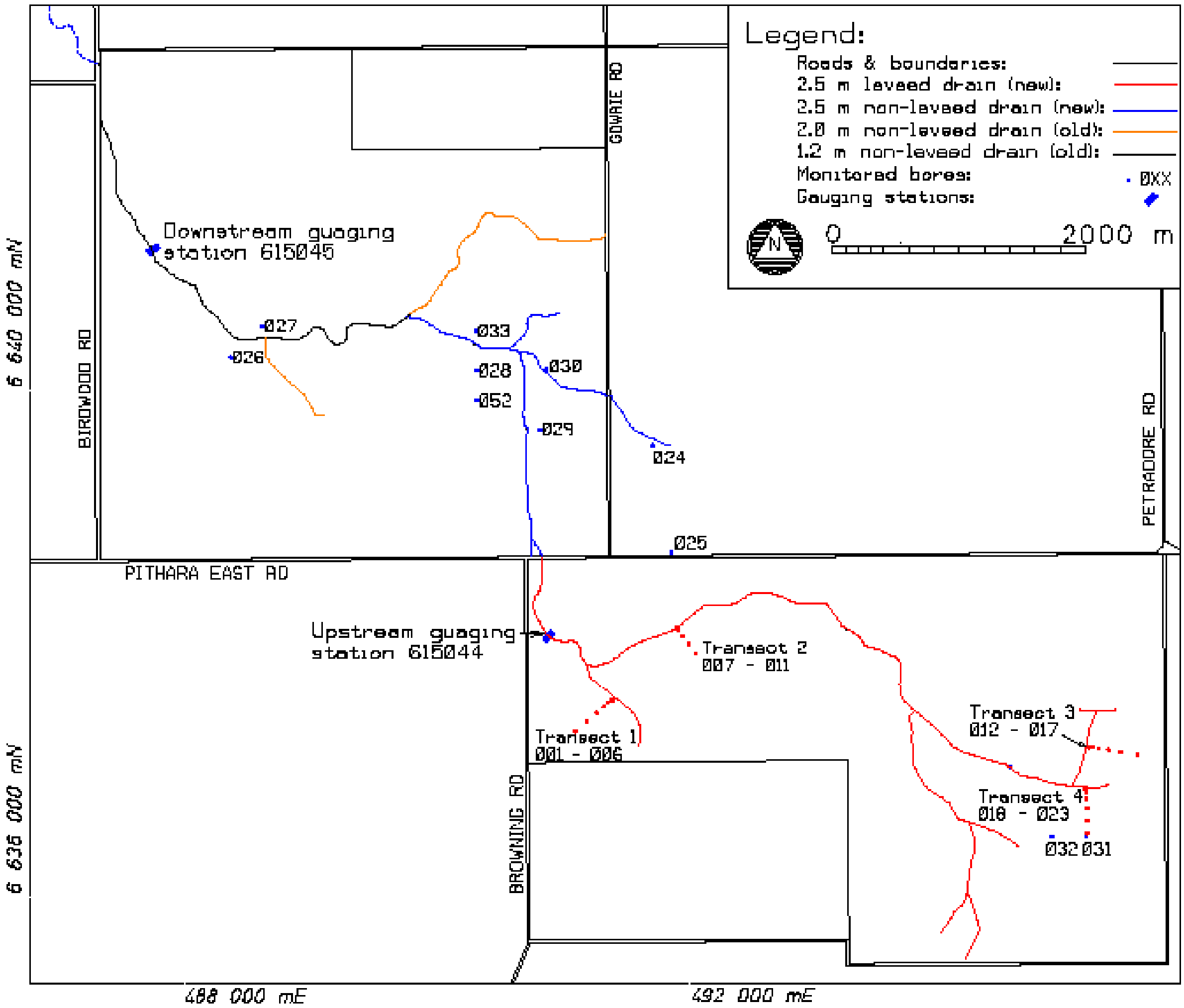

Figure 10 Pithara drainage project site plan.

Figure 11 Excavator at the Pithara drain

3.2 Monitoring and measurements

There was a strong emphasis on monitoring discharges from the drains in response to changes in groundwater levels. This was done so that as much as possible sound ‘cause and effect’ relationships could be established between the two and where appropriate attributed to the drains. The intent was to reinforce the findings from Pithara over and above other drains which produced ambiguous results because watertable changes alone were measured.

Rainfall

An automatic (pluvio) rain gauge (Station 508041) was placed just north of the upstream gauging station so that it was positioned about midway along the drain. Rainfall was recorded at five-minute intervals from 4 June 2004 to 10 January 2007 (Section 2.1).

Water levels

A total of 37 bores were routinely measured as part of the groundwater monitoring program. Over time, the landowners had previously drilled 41 monitoring, test and production bores across the property, nine of which were adopted into the project monitoring program (PT024–030, PT032 & PT033) (Fig. 10).

Although the landowners had installed 41 bores of various depth and construction along the valley floor, no historical water level or water-quality data were available. All except PT032 were north of the Pithara East Road. Bores PT024–027 were designated ‘comparison bores; that is, bores in a position thought to be unaffected by the drainage and so useful to distinguish natural water level changes from drainage-caused changes.

The nine existing bores were first measured in December 2003 and then subsequently as part of the monitoring program mentioned below. The remaining 32 original bores were measured once in December 2003 and some again in May 2004.

Drained watertable changes were measured with four alignments of transects of bores extending perpendicular from one or other side of the upstream leveed drains (Figs 10 & 12). The 27 transect bores (PT001–023) were specifically drilled to measure changes in the watertable profile alongside drains. Transects 1, 3 and 4 each had six shallow bores at 20, 50, 100, 175, 275 and 400 m from the centre of the drain and transect 2 extended to only the 275 m bore. All of the shallow transect bores are about 4 m deep with the lower 3 m fully slotted and gravel packed.

The bores at 20 m were paired with deep bores drilled to basement to measure the deeper groundwater levels and determine the aquifer characteristics beneath the drains (Section 2.5). The deep bores varied from 15 to 30 m with the lower 6 m slotted and gravel packed. The depth and construction of all of the newly drilled project bores were logged and reported at the time of drilling (Dogramaci 2004 & Appendix CD 2.5).

Monitoring the transect bores commenced in May 2004 only two months before drain construction in July. The lack of long-term historical data prevented effective comparison between before- and after-drain groundwater trends and hindered the assessment of any water level response.

Water levels were measured manually fortnightly to monthly from May 2004 to December 2006. Also, automated water level sensors, placed in transect and comparison bores just before drain construction, operated 10 June–12 November 2004. The sensor loggers recorded six hourly to capture any drawdown caused by the newly dug drain.

In-situ salinity and pH were measured in all monitoring bores at 1–3 monthly intervals. A large range of major ions, metals, total nitrogen and phosphorus were measured by laboratory analyses of samples from the deep and shallow bores PT001, PT007, PT012 and PT018, and comparison bores PT025 and PT026 (Section 4.4). Samples were collected September 2004, May and September 2005 and March 2006 to coincide as best as possible with the ends of summer and winter.

Individual bores will be further referred to by number only; that is, PT001 to PT065 become bores 001 to 065.

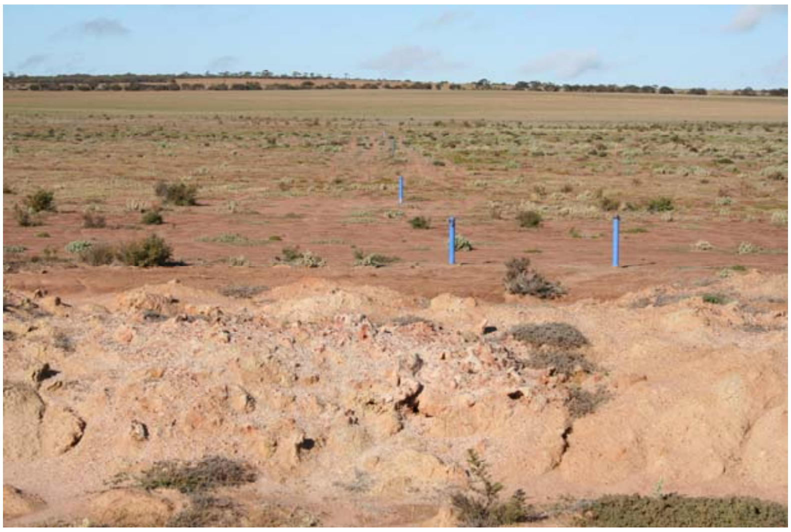

Figure 12 Transect 4 extending from the drain with deep and shallow bores in the foreground

Drain discharge

Discharge was measured at two points along the drain (Fig. 10). As the drain upstream of the upstream gauging station (615044) was leveed, only groundwater inflow and any rainfall runoff from the drain area enclosed by the levees was measured. In contrast, the downstream gauging station (615045) measured groundwater inflow into the whole drainage system as well as surface runoff from its 13 200 ha catchment.

Placed in the modified watercourse 1600 m upstream of the property boundary the downstream drain gauging station measured discharges and salinity from the following:

- 2110 m of the original 2.0 m deep drain
- 2610 m of the 1.2 m deep creek excavation
- 4710 m of new 2.5 m deep drain downstream of the upper gauging station
- 9250 m of new, leveed 2.5 m deep drain upstream of the upstream gauging station
- 13 200 ha of mostly cleared farmland catchment surrounding all of the above drains.

The upstream drain gauging station was placed about 700 m south of the Pithara East Road and measured discharges and salinity from:

- 9250 m of new, leveed 2.5 m deep drain connecting into the downstream drains (as above)
- 11.1 ha surface catchment formed by the drain structure enclosed between the levee banks.

The gauging stations had weir structures built in the drain and float wells alongside (Fig. 13). Interconnecting pipes from upstream of the weir and float well maintained the same water level between the two. A float and transducer mechanism inside the float well measured drain water level at five-minute intervals that was later converted to flows. Stations 615045 and 615044 commenced measuring on 4 June 2004 and 24 June 2004 respectively. Both operated until the project closed on 9 January 2007. In July 2004, the gauging stations were fitted with EC probes that recorded salinity at 15 minute intervals.

In-situ pH and laboratory analyses of drain discharge were done at the same time as the bore samples and for the same range of major ions, metals, total nitrogen and phosphorus.

Figure 13 Downstream gauging station 615045

4 Results

4.1 Groundwater level responses to the drain

Watertables measured along the four bore transects (Section 3.2) showed a noticeable reduction in response to the drains up to 50 m from the channel. From 50 m to 175 m there are signs of possible watertable responses, but these become less clear with increasing distance, intermixed with the natural variability of recharge and groundwater evaporation.

Groundwater responses from the four transects are discussed individually in Appendix A. The following distances represent the span of noticeable to potential reductions in watertable height alongside the drain:

- Transect 1 50–100 m
- Transect 2 50–175 m
- Transect 3 50–100 m
- Transect 4 50–175 m

The watertable falls attributable to the drains were in the order of tenths of a metre within tens of metres of the drain, and sometimes centimetres out to the distances mentioned above. The watertable reductions were unlikely to be enough to lead to any noticeable reduction in topsoil salinity and cereal improvement in cropping conditions alongside the drains.

Deeper groundwater levels within 20 m of the drains appeared to experience delayed and/or smaller reductions in response to the drains. The groundwater heads beneath the bore transects mirrored or were higher than the watertable. The groundwater head is highest beneath transect 3 where it remained consistently 1 m above the post-drained watertable height (Fig. 36).

The deeper and shallow groundwater heads tend to reflect that the underlying aquifer experiences varying degrees of confinement across the valley floor. Confinement is most likely caused by the variable permeability of the weathered profile between the more permeable underlying disintegrated basement (saprock) aquifer and overlying surficial sediments (Section 2.5). The aquifer tends to be semi-confined within the upper parts of the valley floor, tending towards unconfined in the lower and more severely salt-affected parts.

Assessing the effect of the drains from the groundwater hydrographs alone is limited by the small changes measured, the short duration of pre-drain measurements and the onset of winter rainfall at the time of drain construction.

4.2 Groundwater salinity

Groundwater salinities beneath the valley generally increased:

- towards the centre of the valley floor or drainage line
- downstream to the west
- with increasing groundwater depth.

In-situ measured groundwater salinities across the Petrador Farms ranged from 1750 to more than 74 000 mg/L. The average salinity in the deep transect bores was consistently 2000–10 000 mg/L higher than for the average of the shallow bores along the transect, except beneath transect 3 (Fig. 14). Beneath transect 3, the salinity of the deeper groundwater was both comparatively low and stable compared to the other groundwaters, remaining within a range 7000–10 000 mg/L (Appendix CD 4.0b). At this salinity, the groundwater was on average about 15 000 mg/L less than its saline watertable above (Fig. 14).

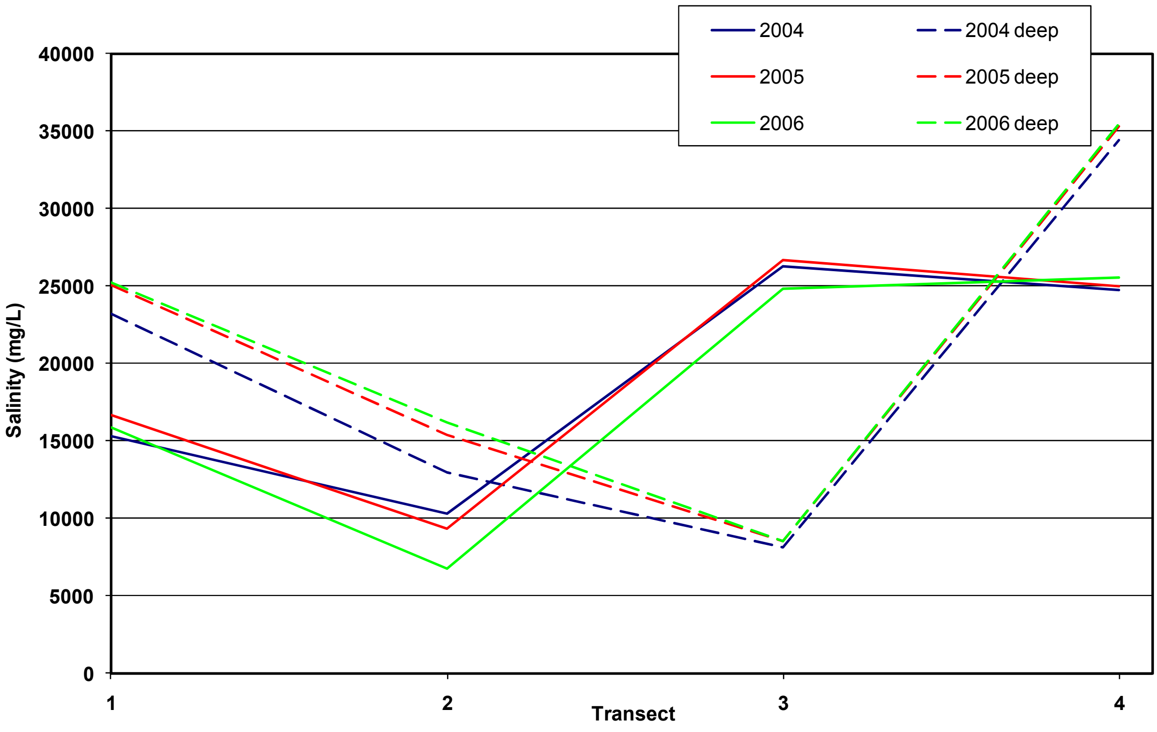

Figure 14 Average annual watertable and deep groundwater salinities (transects 1–4)

Watertable salinities may be influenced by localised recharge, evapoconcentration (Dogramaci et al. 2009) and deeper groundwater rise. As a result, salinities show no consistent seasonal trends and, in most cases, bores in the same transect responded differently (Appendix CD 4.0b). Transect 1 produced the most consistent salinity changes corresponding with variations in climate and watertable levels. Most of the salinities in transect 1 rise during the autumn months (March–May) (Fig. 15) associated with falling watertables (Appendix A) which is probably in response to groundwater evaporation.

Watertable salinities at the outer ends of the transects were lower than and more uniform than those closer to the centre of the valley floor or drain alignment (Figs 15 & 16). The watertable is usually deeper at the outer end of the transect, producing an association between greater depth to watertable and lower and/or more uniform salinities. With no deep bores close to the footslopes it is not known if deep groundwater there was less saline than closer to the centre of the valley.

Except for transect 3, where deep groundwater salinities were the lowest, salinities tended to be in the middle to upper range of salinities of their respective transect watertable bores (Fig. 15). The salinities also tended to be stable, mostly fluctuating within a 5000 mg/L range.

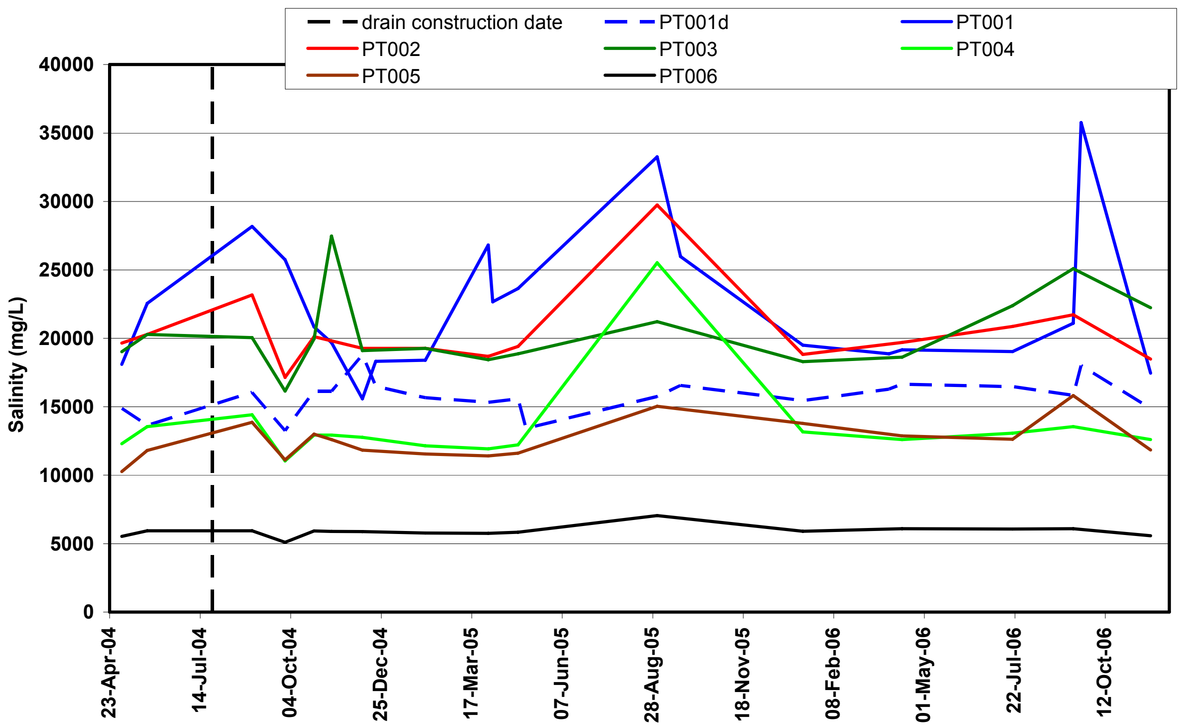

Figure 15 Groundwater and watertable salinities beneath transect 1

The average watertable salinity along the transects increased towards the middle of the valley (Fig. 16). The watertable salinity profiles and deep groundwater salinities reflect the groundwater discharge and salt export patterns across the valley floor. Towards the centre of the valley floor where deeper groundwater heads are highest and watertables shallowest, upward leakage and evapoconcentration of deep groundwater causes increasing watertable salinities during the summer and autumn months. Towards the end of winter watertable salinities fall when shallow groundwater is diluted with rain-fed recharge.

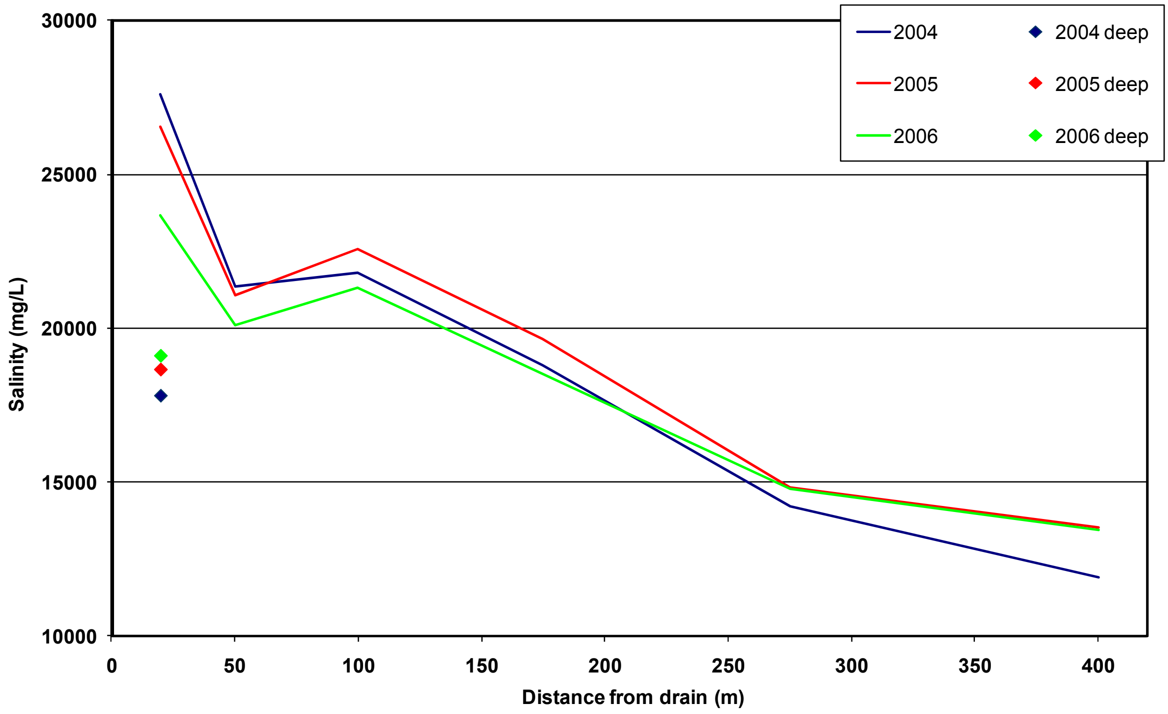

Figure 16 Average annual watertable salinity profiles and deep groundwater salinities (transects 1–4)

Lower salinities at the periphery of the valley floor are probably the result of the greater depth to the watertable providing less opportunity for the evapoconcentration of shallow groundwater. Fresh hillside seepage may also contribute to dilution of saline water (Section 2.6) in this area. The high local variability in watertable salinities provides an indication of the absence of horizontal groundwater movement because the soil is fairly impermeable and the valley floor is flat.

Irrespective of watertable level reductions drain responses could still include adjacent reductions in watertable salinities. This might happen if the drain depressurises the deeper groundwater reducing the rate of upward leakage and/or enables salts to be leached from the shallow watertable (Chandler & Coles 2003).

In general, watertable salinities within 100 m of the drain decreased steadily through the monitoring period, while those at 100–400 m and the deeper groundwaters remained unchanged or rose slightly. By 2006, watertable salinities had fallen to below their 2004–05 average levels within 100 m, while those beyond were stable or had increased (Fig. 16). Deeper groundwater average salinities rose each year through 2004–06, rising by just over 1000 mg/L. The salinities of the comparison bores 024, 26 and 027 rose by about 5000 mg/L during the monitoring period (Appendix CD 4.0b).

There was no immediate effect of the drain on reducing watertable or deeper groundwater salinities. The measurements were probably too infrequent to detect any rapid salinity changes if they did occur. The 2004–06 average reduction in watertable salinity of up to 4000 mg/L, close to the drain, conforms with expectations of salinity reductions in response to drainage. Ongoing monitoring is needed however to confirm that these and any ongoing salinity reductions are in response to the drain and not just a natural groundwater response to a succession of dry years.

4.3 Groundwater pH

The watertable pH ranged from 2.6 to 8.8, although the lowest in the transect bores was 3.5. Very low pH levels (below 3.0) were in bore 027 located towards the downstream more severely salt-affected end of the drain. The groundwater pH generally increases along the valley floor towards the drain outlet, corresponding to the pH of the drains (Sections 4.5 & 4.6).

Contrary to this, the pH along the upstream valley floor tended to be more acidic towards the upper end of the drain in transects 3 and 4 (Fig. 17). This perhaps reflects a compartmentalised groundwater system with the upstream aquifer disconnected from the downstream aquifer. It could also reflect different geochemical processes affecting the upstream groundwater system.

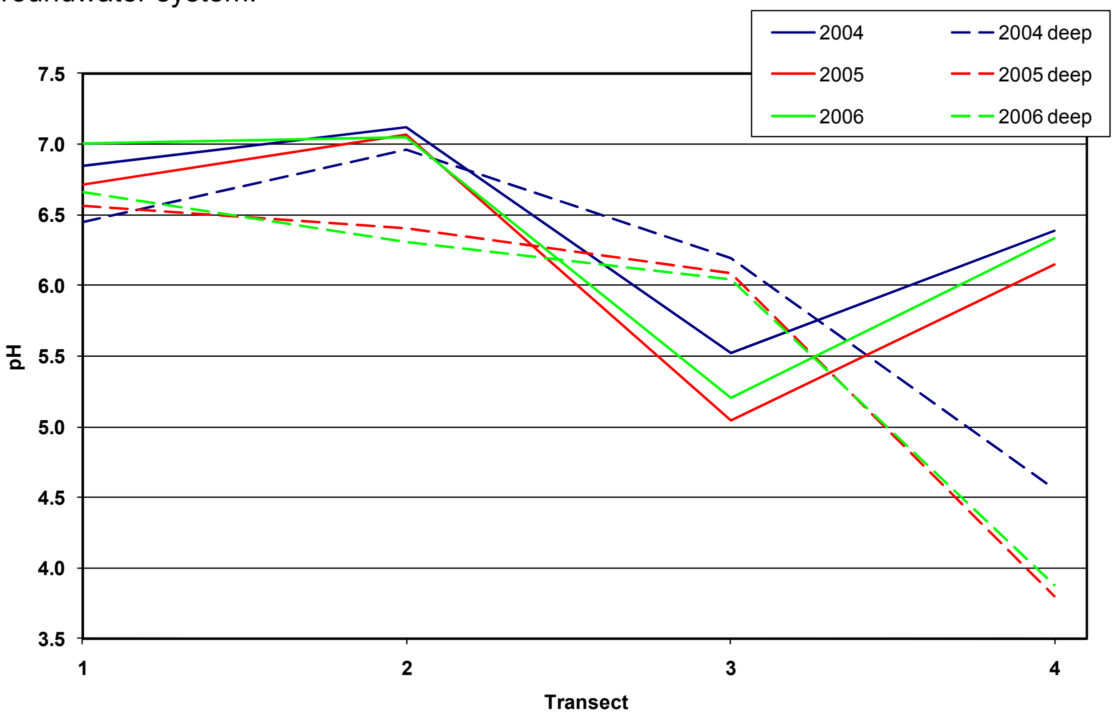

Figure 17 Annual pH averages for the watertable and deep bores of each transect

The average watertable pH increased beyond 100 m from the drain alignment (Fig. 18). There appears a relationship between increasing salinity and decreasing pH from a comparison of Figures 14 and 17 and 15 and 18. This is contrary to the normally accepted relationship between high levels of salinity (sodium ions) and increasing alkalinity (Ritzema 1994).

In all but transect 3, pH in the deep bores was usually lower than in the watertable bores (Appendix CD 4.0b). Deep groundwater pH was 3.8–7.0 throughout the monitoring period. Within transect 3 the groundwater pH was consistently higher than watertable pH in all its shallow bores except 017.

The low pH is mainly due to high concentrations of dissolved iron reacting with oxygen in infiltrating waters (Shand & Degens 2008). The reaction results in the precipitation of iron, forming iron oxide, and, if no neutralising agents are present, lowering the pH. Because dissolved iron is mobilised and precipitated in much the same way as other salts, high levels of dissolved salts and iron go hand in hand beneath the valley floor. The relationship between higher levels of salts and higher levels of iron is the foundation of the relationship between salinity and pH at this site.

The iron concentrations and geochemical processes in the groundwaters affected the pH of in each bore independently. In-situ groundwater pH in any one bore varied by as much as 3 pH units in response to different sampling techniques and the chemical reactions they might cause. Pre-sample bailing of any one bore could result in pH rising or falling in comparison with the pH of the standing water. Because the groundwater yield of some bores is low, uniform sampling techniques could not be applied. As a result, the pH measured in some bores may be strongly influenced by the sampling techniques used.

The groundwater pH showed no detectable seasonal trends or consistent responses to climatic events (Appendix CD 4.0b); perhaps from too few measurements to capture possible trends, or the absence of trends because groundwater pH is controlled by local geochemical processes.

In light of the groundwater responses, the only possible and expected watertable pH response to the drain was increasing acidification within 50 m (Fig. 18). This could come about either in response to deeper groundwater rise or accelerated iron oxidation in the shallow groundwater.

Lowering the watertable close to the drain can increase the head difference between the watertable and deeper groundwater enhancing the upward leakage of acidic groundwater. This upward leakage of more acidic or iron-rich groundwater can in turn increase acidity of the shallow groundwater, resulting in the falling average pH results close to the drain (Fig. 18). Alternatively, or in combination, draining the watertable or adding iron-rich groundwater increases the potential for iron oxidation and subsequent falling pH.

Like salinity reduction close to the drain (Fig. 16) the 0.5 unit pH reduction could be a response to below average rainfalls for 2005–06. Below-average rainfall combined with continuing groundwater evaporative loss could produce the same pH-lowering effect on the watertable as drainage. Ongoing monitoring is needed to sort out drain and/or weather effects on groundwater pH.

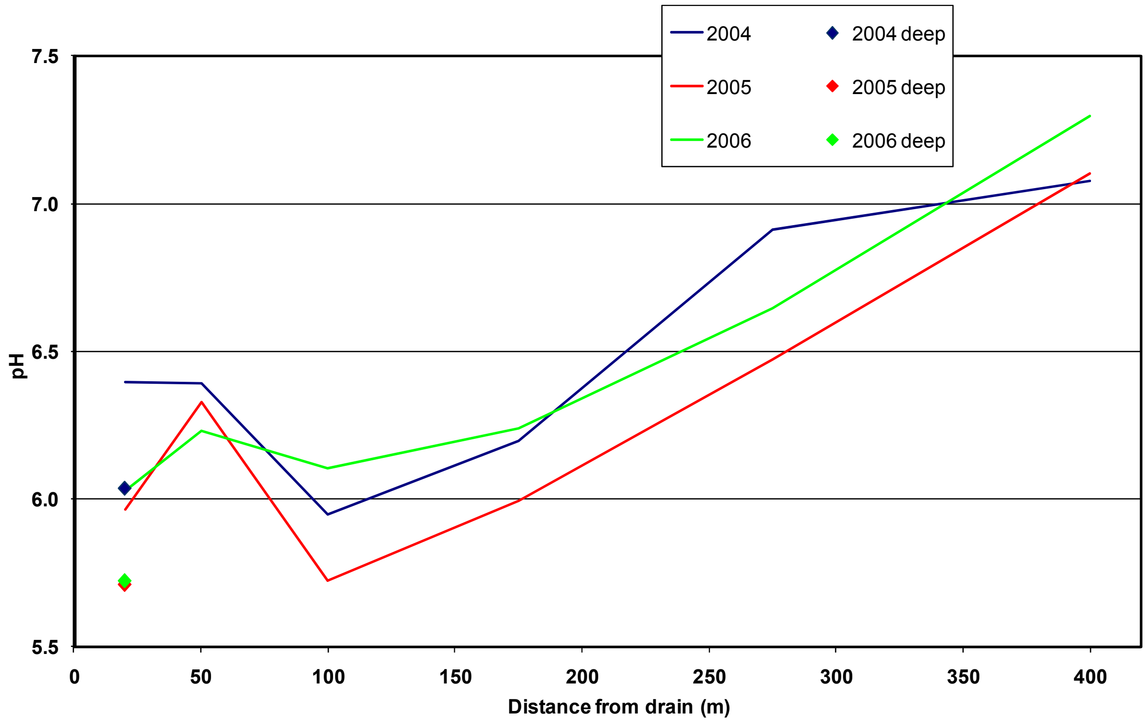

Figure 18 Yearly pH average for the watertable and deep bores along the transects

4.4 Groundwater chemistry

The groundwaters have similar major ion compositions to sodium chloride dominated seawater, like most groundwater and surface water in the south-west of Western Australia (Mayer et al. 2005). This suggests that the most important groundwater chemistry process is the concentration of salts derived from marine aerosols, transported to groundwater by rainfall recharge and concentrated by evapotranspiration (Dogramaci & Yesertener 2001).

Although the major iron compositions are similar to seawater, the ion concentrations vary, probably influenced by local physical and geochemical processes (Appendix CD 4.4). Concentrations of chloride may double for example between one sample and the next three months later. But during that period, concentrations of other ions may increase by only a small percentage, or even decrease.

It is beyond the scope of this report to investigate what appear to be complex and possibly somewhat random geochemical interactions of the groundwaters. The important point is to show the origin of the salts and some groundwater properties that might influence drain discharge quality.

Low pH groundwaters contain high levels of dissolved metals (Shand & Degens 2008). The particularly low carbonate in 026 coincided with this bore's often low pH (3.5). The low pH in this bore is accompanied by the highest concentration of almost all the metals, including 250 mg/L of dissolved iron on 2 September 2004, and concentrations of aluminium, copper and lead that were well above those of the other bores.

Dissolved metals originate from the regolith and levels can vary significantly in response to various regolith compositions. The dissolved metals (selenium, mercury and cadmium) that cause concern when discharged by drainage into the natural environment were below detection limits at Pithara.

Total phosphate and nitrate concentrations were mostly acceptable for livestock drinking and ecosystem protection (Environmental Protection and Heritage Council 2000). Like salinity, pH and major ions, these nutrient concentrations vary with soil conditions, land use and over time.

Watertable phosphate as total phosphate concentration fluctuated in any one bore (0.18–3.4 mg/L) with most around 1.5 mg/L. Bore 007 was the exception with 11–21 mg/L. These high concentrations, not reflected in the deep groundwater, are likely to result from local point source contamination such as fertiliser spillage.

Bore 007 does not contain correspondingly high levels of total nitrogen (N); levels were well within the range of the other bores – from below detection to 0.18 mg/L.

Lower concentrations of total P in deep bores than in their watertable pair appear to be the only pattern in the analysed groundwater nutrient concentrations.

4.5 Upstream drain discharge

The drain upstream of station 615044 discharged 103 800 kL during the 921-day monitoring period 24 June 2004–31 December 2006. The average discharge from the 9250 m of drain equated to 12.2 kL/d (0.14 L/s) per kilometre length.

Drain discharge consists of seasonally varying volumes of baseflow punctuated by short periods of increased discharge in response to rainfall, and no flow due to high evaporative losses. Baseflow discharge was highest in late July 2004 (up to about 500 kL/d) coinciding with drain completion and continuing winter rainfall (Fig. 19). During the winters of 2005 and 2006 baseflow discharge was steady at around 100 and 60 kL/d, respectively.

Drain flow stopped for several days in January 2005, the first dry summer period after construction. It also stopped flowing intermittently in December 2006 following the fairly dry winter of that year. Baseflow increased from near zero, to 170, and then 250 kL/d in response to 140 mm summer rainfall in January–February 2006. By March, baseflows had fallen to about 70 kL/d, and remained at that rate until the onset of winter. The baseflow appeared to be sustained through the summer to early winter months in 2006 by the preceding intense January–February rainfall.

Evaporation had a noticeable effect of baseflow rates. As flows decreased it was first noticed that baseflows stopped during the hotter part of the day, and recommenced during the cooler parts and evenings. During prolonged hot and dry periods, baseflows ceased completely, but could re-commence in the absence of rainfall with just cooler weather. About 60 000 kL of drain flow was lost to evaporation (Section 5.2).

Daily drain discharges varied by an order of magnitude in response to water from rain falling directly into the channel and runoff from the drain structure. The 676 mm of rain that fell onto the 11.1 ha drain structure (Section 3.2) was equivalent to 75 000 kL, although only about 10% (7200 kL) became runoff and contributed to the measured discharge (Section 5.2).

Increased drain flows were caused by pumping into the channel from a nearby groundwater pumping scheme from 18 February–14 April 2005 and 11 January–10 February 2006. Groundwater with a salinity of 6100 mg/L was irregularly pumped at a rate of 2 L/s into the drain channel at transect 4 (Fig. 10). During 2005 and 2006 about 2190 kL and 3200 kL respectively of groundwater was pumped into the drain.

It was thought that drain discharge would have stopped during more days in February and/or March 2005 but for the groundwater pumping contributions. In 2005, the groundwater pumped into the drain made up 59% of its discharge, reducing to 37% in 2006. Although a larger volume of groundwater was pumped into the drain in 2006, it was proportionally less than the volume of drain discharge for the period.

Drain flow salinities fluctuated mainly due to changes in water volume caused by evaporation, rainfall and runoff. The average daily flow-weighted salinity was about 49 000 mg/L, ranging from 8000 to 155 000 mg/L (Fig. 20). The exceptionally low flow-weighted salinity (about 2000 mg/L) was caused by surface water runoff entering the drain during construction on 8 July 2004. Salinities rose above 155 000 mg/L in response to the evapoconcentration of ponded water in the channel. The total flow-weighted salinity for the entire monitoring period was 31 500 mg/L; more saline than most of the underlying groundwaters and adjacent watertable (Fig. 14).

High salinities without drain flow resulted in zero flow-weighted salinities such as in January 2005 and December 2006 (Fig. 19). When low flows stop and restart salinities fluctuate between zero and very high levels.

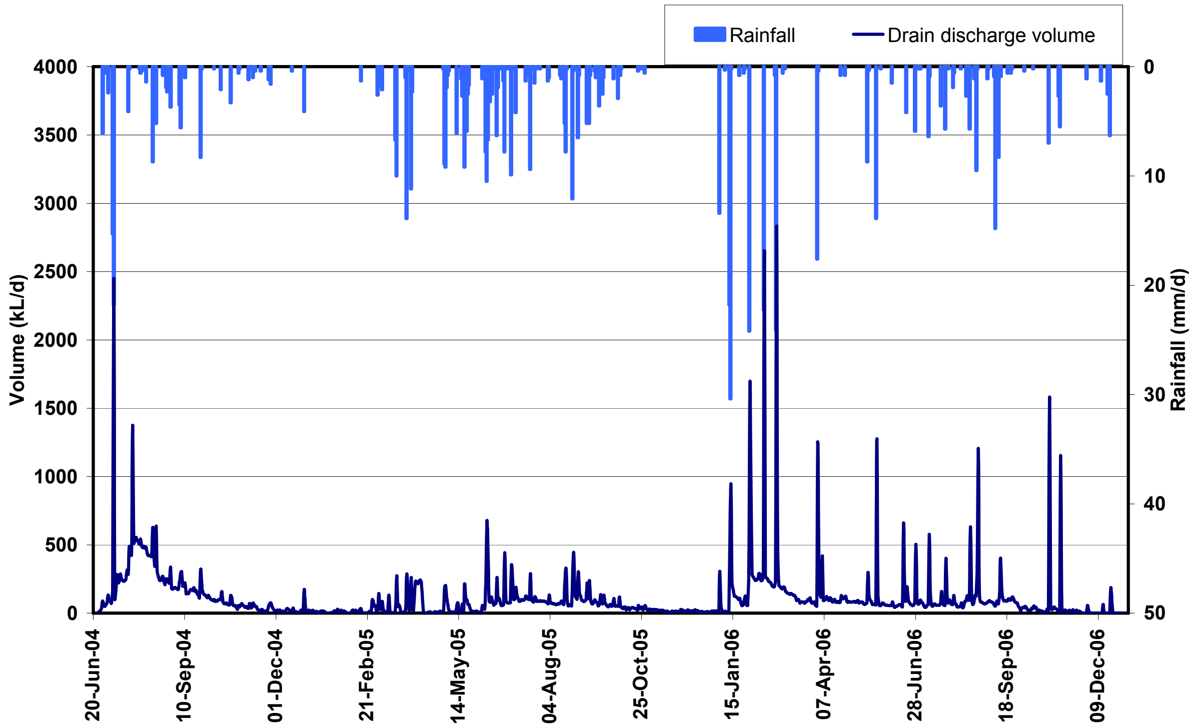

Figure 19 Daily upstream drain discharge and rainfall

The onset of rainfall and runoff diluted saline water in the channel. The lowest post-construction drain flow salinity (7850 mg/L) was in response to fresh water from rainfall and runoff on 13 January 2006 (Fig. 20). The magnitude of each dilution event depended on the proportions of saline drain flow to rainfall and runoff. During low flows, even small contributions from rainfall, evident by the spikes in the hydrograph, sometimes caused large salinity reductions. Conversely, quite substantial winter rainfall usually reduced salinity slightly due to high baseflows at those times.

Flow-weighted salinity fell from 145 000 mg/L to about 15 000 mg/L in response to the groundwater pumping in February 2005 (Fig. 20). During this time pumped inflow represented the greater proportion of the drain flow, hence the large reduction in discharge salinities. Flow-weighted pumping-affected salinities were about 25 000 mg/L between the January–February 2006 rainfall events. The higher 2006 than 2005 salinities were caused by the lower proportion of pumped groundwater to drain flows despite more groundwater pumped into the drain in 2006.

There was no apparent significant change in the salinity trend of the drain discharges beyond those directly influenced by the pumping and the climatic variables of rainfall and evaporation. Baseflow salinities returned to about 33 000 mg/L around August in each of 2004, 2005 and 2006 (Fig. 20). During these mid winter months, the groundwater inflows that contributed to baseflows were at their most stable with salinities least influenced by evaporation.

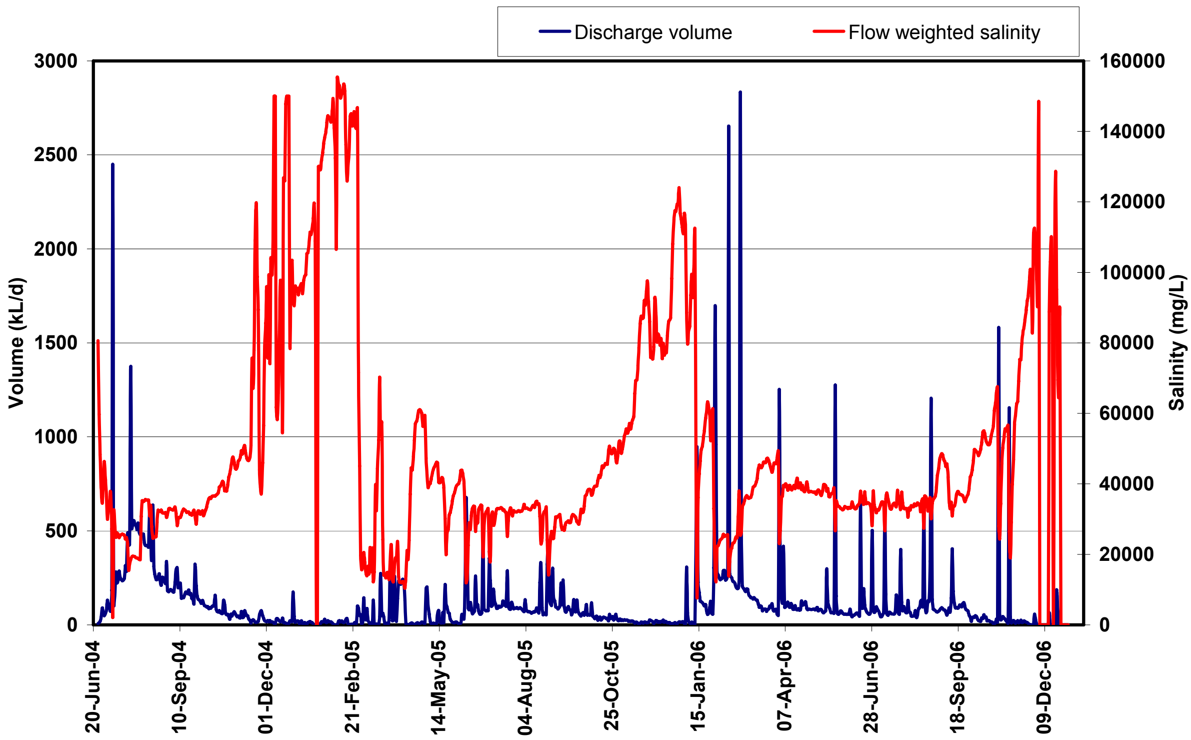

Figure 20 Upstream drain discharge and flow-weighted salinity

Drain discharge exported 3270 t of salt. Most of this was from groundwater inflow to the drain, as other sources such as surface water runoff from surrounding farmland were excluded. The average salt flow equated to 0.38 t/d/km, ranging 0–7.8 t/d/km. Daily salt loads mostly mirrored drain discharges except on occasions during and immediately following periods of peak rain-fed discharge.

Rainfall following extended dry periods appeared to wash accumulated salts from the channel batters and floor, causing greater than expected increases in daily salt loads (Appendix CD 4.1). Rainfall causing large increases in drain discharge temporarily displaced ponded saline water and salt loads. Salt loads sometimes remained low for several days afterwards until the fresh water was again displaced by the saline groundwater inflows.

The major ion composition of the upstream drain flow very closely resembles that of the underlying groundwaters. There was an almost equal proportional increase in the major ion concentrations in the groundwaters and drain flows, reflecting the evapoconcentration of the groundwater within the drain (Fig. 21). The average concentration of the major ions was 56 000 mg/L compared with 35 000 mg/L for seawater (Appendix CD 4.4).

Although the average major ion concentration of the drain flow was 1.6 times more than seawater, this was mostly due to the very high concentrations measured in February 2005. Excluding February sample results, major ion concentrations in average drain flow and seawater correspond more closely except for calcium and magnesium levels which were low. The low calcium levels, at about 50% of seawater, were unexpectedly low given the abundance of calcium carbonate nodules in the upper soil profile along some sections of the drain (Section 2.4).

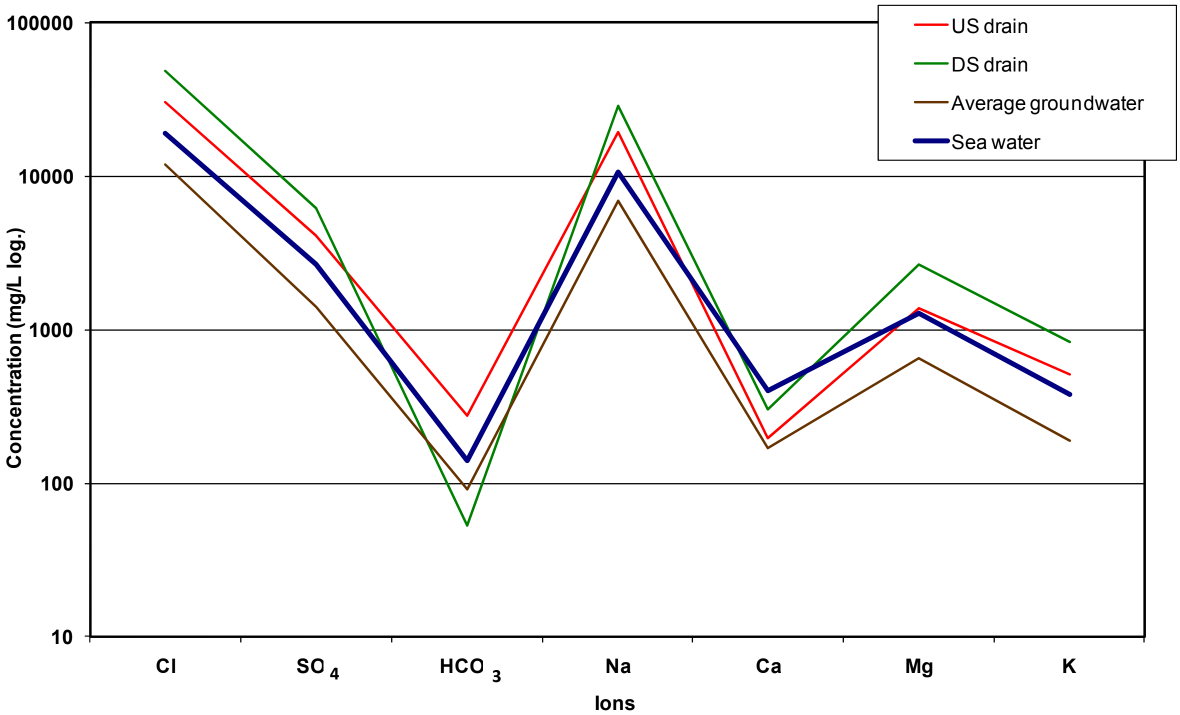

Figure 21 Drain flow major ion concentrations compared to groundwater and seawater

The laboratory-measured pH values indicate neutral waters unlike the acidic waters in many other drains in the central and southern Wheatbelt. Many drains, including the downstream drain, produce acidic waters (Silberstein et al. 2005) with high concentrations of dissolved metals, such as iron, aluminium, as well as trace metals and rare earths. The concentrations of these metals in the upstream drain were extremely low and often below detection limits (Appendix CD 4.0c), probably because most of the trace metals and rare earths are insoluble at pH >5.

The 30 in-situ discharge pH measurements were 6.12–8.63 (Fig. 22). Although there are too few measurements to identify the possible full range of pH, it was evident that low pH was associated with lower drain discharge rates. As the low discharge rates are dominated by groundwater inflow, this confirmed groundwater inflow as the primary source of the low pH drain discharge.

Although pH changed quickly in response to changes in the sources of inflows (groundwater, rainfall and/or runoff) most pH values were 7.5–8.5 (Fig. 22). The pH showed a slight upward trend mostly between 7.5 and 8.0 in 2004 to 8.0 and 8.5 in 2006. Although not demonstrated, the upward trend was thought to be caused by the increasing pH-buffering capacity of carbonate-rich topsoils as they were eroded from the upper parts of the drain structure and accumulated in the channel.

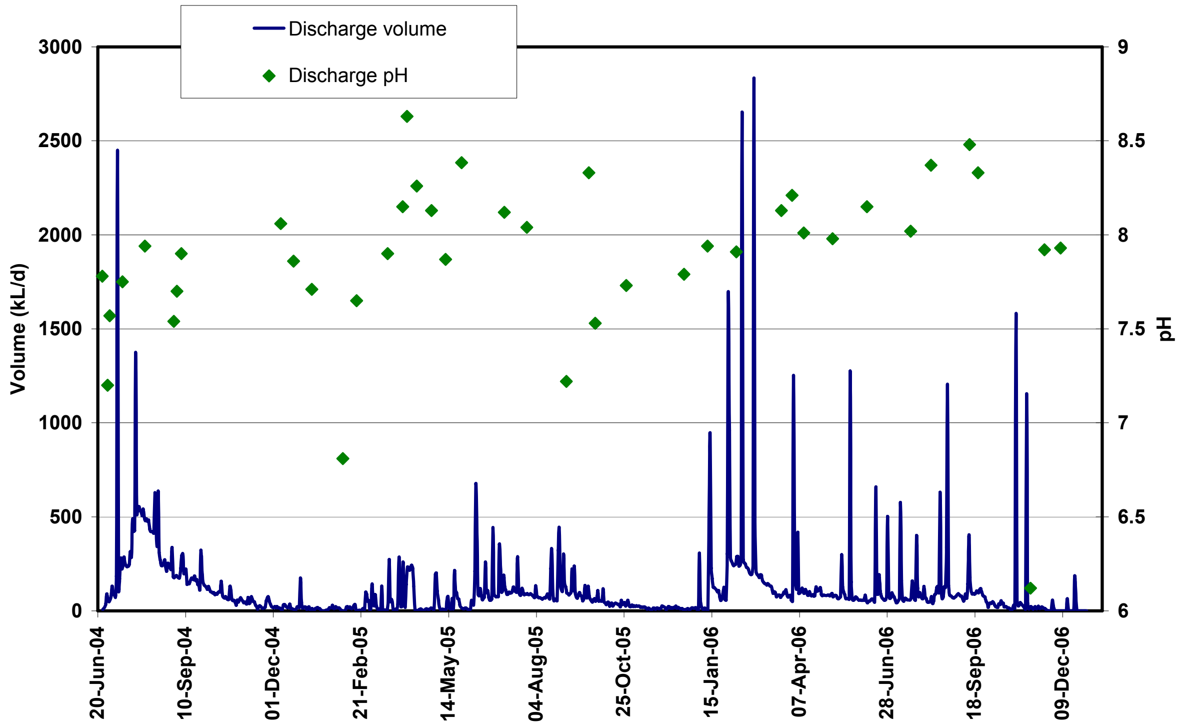

Figure 22 Upstream discharge and pH

4.6 Downstream drain discharge

Between 24 June 2004 and 31 December 2006, 306 600 kL was discharged through the downstream gauging station 615045. This was an average discharge of 17.8 kL/d/km length of drain at an average rate of about 0.2 L/s/km over the 921 days. The downstream discharge hydrograph is almost identical to that of the upstream drain, but with most daily discharge being two to two and a half times greater.

July 2004 post-drain construction baseflows of 1400 kL/d fell to around 225 kL/d in the winter of 2005 and 140 kL/d in 2006 (Fig. 23). Low summer baseflows from the drain were sustained mostly by local groundwater inflow from beneath the severely salt-affected land immediately upstream of and surrounding the gauging station. Although summer baseflows often fell to only several kilolitres per day these were sufficient to maintain drain discharges.

Reduced groundwater contributions to the downstream creek section (Section 3.2) caused the greater than expected evaporative loss of discharge from the upstream drains. This often reduced or stopped their discharges being measured at the downstream gauging station. There were some days during summer when small discharges from the upstream gauging station evaporated before reaching the downstream gauging station. There were also some days when the upstream drain had stopped flowing and only discharge from the downstream drain was being measured at the gauging station.

Subtracting the total upstream volume from downstream discharge volume gives about 202 800 kL of discharge from the downstream drain alone. If re-calculated from daily discharges the volume increases to 204 500 kL, with the addition of 1700 kL measured at the downstream gauging station during days when there was no flow from the upstream drain. The average discharge from the 9.4 km length of downstream drain alone was about 23.6 kL/d/km (0.3 L/s).

Although rainfall increased discharges in the downstream drain more than in the upstream drain runoff did not seem to frequently enter the drain from the surrounding catchment. Direct comparison of the downstream and upstream drain discharge hydrographs (Figs 19 & 23) suggests that noticeable volumes of surface runoff entered the downstream drain only twice The discharge from the downstream drain was more than 6 times (15 300 kL) that from the upstream drain on 8 July 2004 in response to 21.8 mm of rainfall. The 17 600 kL of discharge on 13 January 2006 was more than 18 times that of the upstream drain, in response to 30.4 mm of rainfall.

The remaining rainfall events produced downstream drain discharges of about 2.5 times those from upstream. After subtracting the upstream discharges the corresponding peak daily discharges from the downstream drain were about 1.5 times those from upstream. The slightly higher rain-fed runoff contributions could easily be accounted for by the increased potential for runoff provided by the wider channel of the excavated creek bed (Section 3.1).

The contribution to drain flow from the groundwater pumping discharge is still apparent from 18 February to 14 April 2005, but is more difficult to isolate and quantify downstream 11 km from its source.

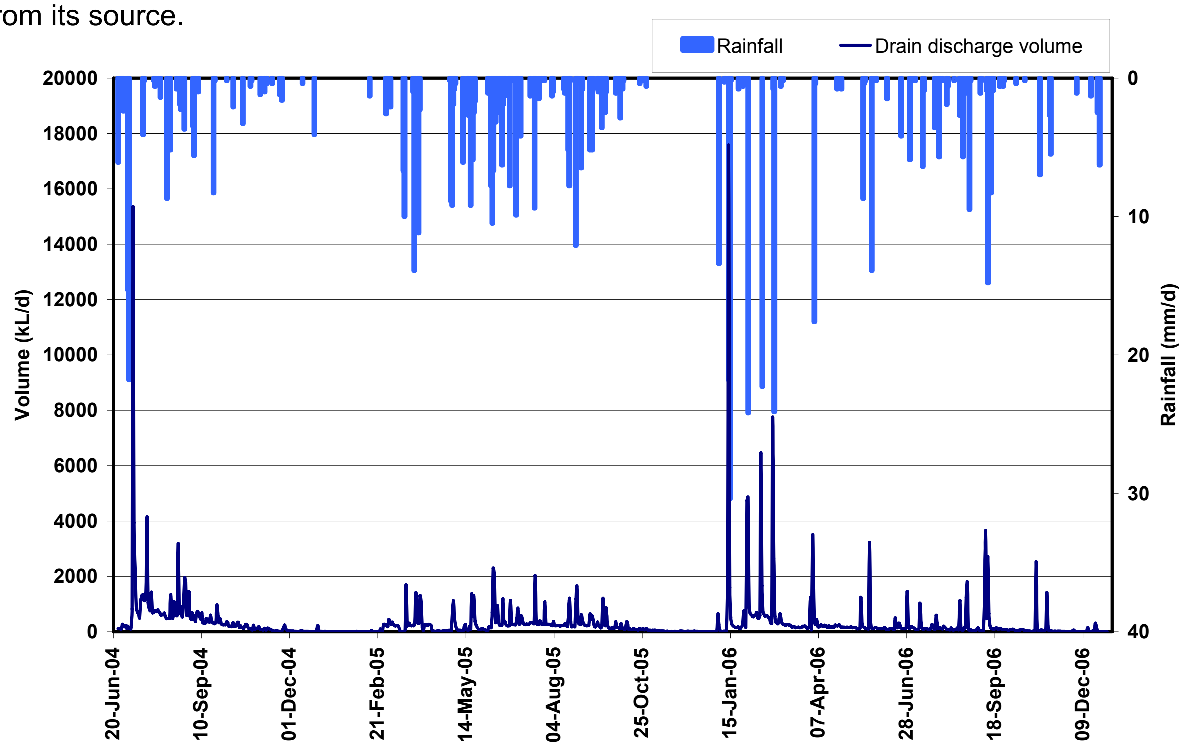

Figure 23 Downstream daily discharge and rainfall

Drain discharge salinity associated with low flows sometimes exceeded the 200 000 mg/L upper limit of the instrumentation (Fig. 24). Large deposits of crystallised salts were sometimes seen at the discharge end and downstream from the drain. The high salinities were caused by the same evaporative processes as for the upstream drain combined with the much higher groundwater salinities (Section 4.2). The downstream drain had a far greater evapoconcentration potential because of the higher evaporative loss from the creek bed between the outlets of the groundwater drains and gauging station.

The average of the daily flow-weighted salinities was about 66 600 mg/L for the measured drain discharges, and 74 800 mg/L for just the downstream drain discharge. Salinities reduced to as low as 10 000 mg/L when diluted by rainfall and runoff, and rose to beyond the upper limit of measurement for the instrument (Fig. 24). The average flow-weighted salinity for the total drain discharge was 35 500 mg/L. In the absence of high evaporative losses, rainfall and runoff, baseflow salinities tended to stabilise at around 45 000 mg/L.

Salt loads from the drain had the same discharge characteristics as the upstream drain, mirroring drain discharge rates. The total salt load from the drain was about 10 900 t, of which 7750 t originated from the downstream section. Because the drain was open to surface water a small proportion of this salt could have originated from surface runoff. Runoff tended to dilute the saline drain flow which suggested a very low contribution of salts from runoff compared to the contribution from groundwater inflow.

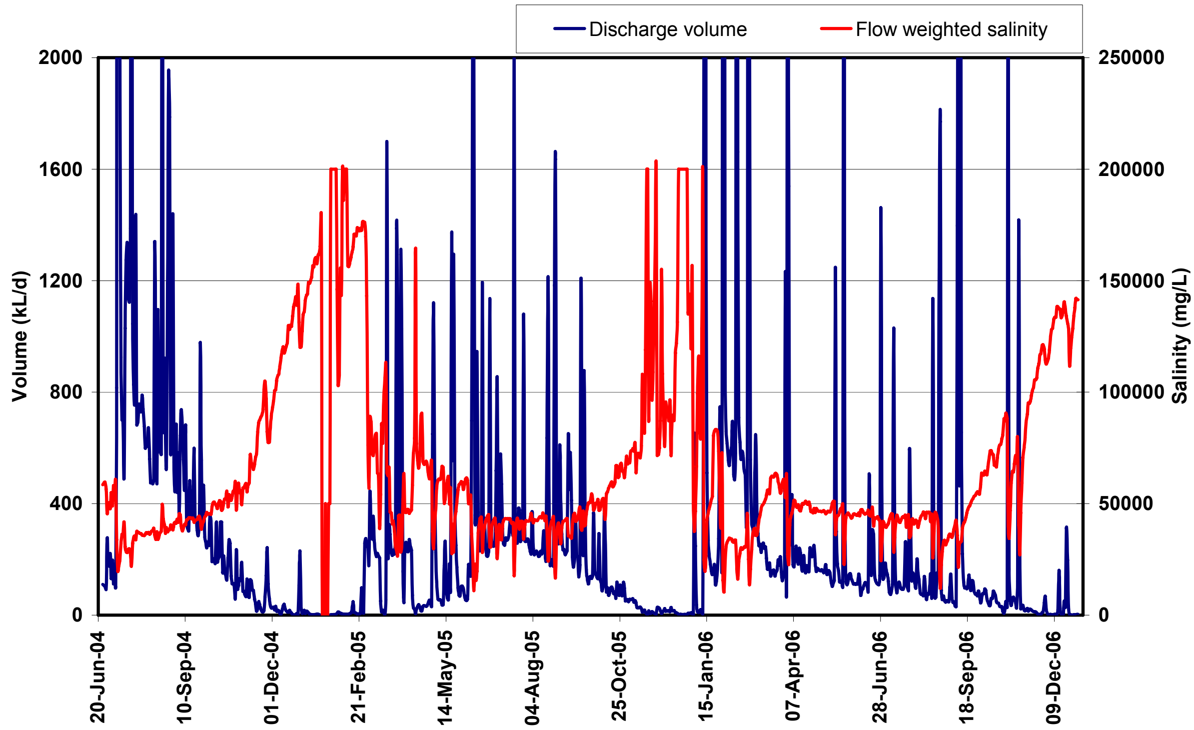

Figure 24 Discharge and flow weighted discharge salinity of the downstream drain

The discharge pH (2.88–8.25) responded to the changing sources of inflow. The drain pH and chemistry (as follows) relate to the combined discharge of the upstream and downstream drains measured at the downstream gauging station. Drain discharge with surface water runoff tended towards neutral–alkaline while discharge dominated by groundwater inflow was acidic. This is reflected in the high discharge–high pH and low discharge–low pH relationships (Fig. 25).

Very low pH groundwater originated mainly from beneath the more severely salt-affected and bare scalded land immediately upstream of the gauging station (Fig. 9). During low flows, drain flow is dominated by iron-rich groundwater from beneath this land, often staining the drain floor orange (cover picture).

The low pH water can dissolve metals and mobilise ions from the regolith. The concentrations of dissolved metals, particularly iron, aluminium, lead and nickel in downstream discharge were around 100 times those in the upstream discharge. Selenium, arsenic and cadmium levels were low to below detection.

Laboratory analysis confirmed salinity closely resembled seawater composition (Fig. 21) with major ions dominated by sodium and chloride and a low carbonate level reflecting the lower pH (Fig. 21). The average salinity concentration from the six samples was 87 444 mg/L – more than 2.5 times seawater, though the concentration falls to 34 900 mg/L when the high ion concentrations measured on 17 February 2005 are excluded. This new average concentration is closer both to seawater and the 35 500 mg/L average flow-weighted salinity discussed above.

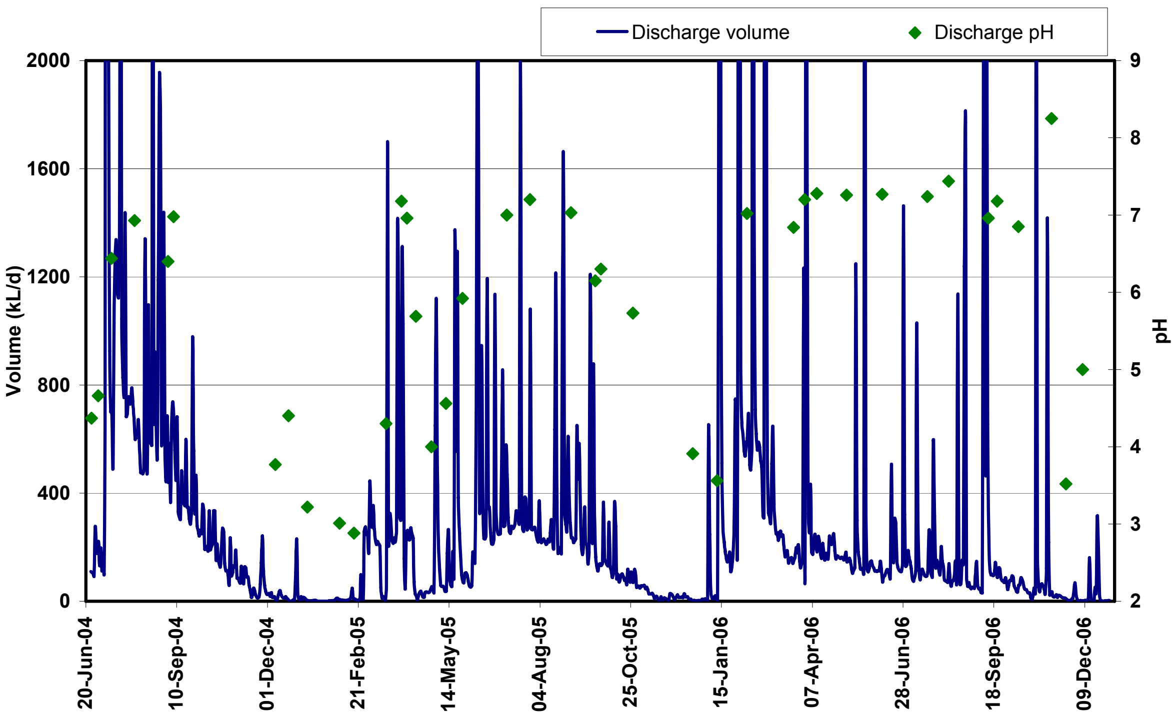

Figure 25 Downstream discharge and pH

5 Water and salt balances

Quantifying the effect of the Pithara drains on the surrounding groundwaters is complicated by its ‘open’ channels and the unbounded design of the drains. Traditional agricultural drainage uses slotted pipes or other permeable conduits buried to below the watertable. The groundwater that seeps into the pipes is the same as the discharges at their outlets. When used for watertable control the pipes are placed parallel to each other in ‘bounded’ schemes that enable the effect of each pipe on its surrounding groundwaters to be easily quantified (Appendix B.1).

The discharges from the Pithara drains were sometimes highly modified by climatic and human variables and at times bore little resemblance to the groundwaters that entered the channel (Sections 4.5 & 4.6). The unbounded design of the drains (Appendix B.1) complicated the assessment of both groundwater drainage and watertable drawdown.

Before the drain, groundwater beneath the saline valley floor was at equilibrium and about a metre below the land surface (Appendix A). The equilibrium was between the inputs to the valley floor aquifer from in-situ recharge and hillside aquifer discharge and its outputs from capillary discharge and possibly some seepage. Groundwater flow along the aquifer is viewed as insignificant in response to its very low gradient and overall transmissivity.

Construction of the Pithara drain affected the output side of this equilibrium or balance by enhancing seepage, otherwise referred to as groundwater outflow from the valley floor aquifer. This groundwater outflow is the groundwater that seeps or inflows into the drain channel, the sources of which are expressed in the water balance (Chandler & Coles 2003):

$$Q = R + (S_{IN} – S_{OUT}) + (Z_{IN} – Z_{OUT})$$

The groundwater inflows to the drain (Q) is shown as sourced from in-situ recharge (R) and the net balance of shallow seepage (S) and deeper groundwater rise (Z) into and out of its groundwater catchment. To be able to solve this water balance requires defining both the groundwater catchment area of the drain and the sources of shallow and deeper groundwaters.

This water balance also assumes that the drain has completely replaced capillary discharge as the mechanism of groundwater output from the catchment. However, the absence of significant post-drain watertable declines at Pithara (Appendix A) suggest that groundwater outflows from the drains were not sufficient to have completely substituted for capillary discharge.

To solve the water balances for the open and unbounded Pithara drains required using two water balance equations. The first for the drain channel separated the relative volumes of rainfall, runoff and evaporative losses that masked the groundwater inflow component of the drain discharges. The groundwater inflows to the channel are the equivalent of ‘Q’ in the above water balance and are the true indicator of drainage efficiency rather than discharge.

The second water balance for the drain groundwater catchment uses a similar approach to the water balance above. The most subjective attribute of the groundwater catchment water balance for the unbounded Pithara drain was estimating the potential extent of its groundwater catchment, further referred to as the drainage site (Section 5.1). Once the area of the site is known, its water balance is conceptualised as a bucket into which go recharge and hillside aquifer discharge, and out of which come capillary discharge and groundwater outflow into the drain. Aquifer discharge is the combination of shallow seepage and deeper groundwater rise into the site, simplified from the water balance above.

The water balances are presented in the logical order of the site first draining into the channel, although this was not the order in which they were solved. In practice, the water balance of the channel was solved first to calculate the groundwater outflow (Q) from the site into the channel. The methodologies and assumptions behind the water balances are provided in Appendix B. Sufficient measurements were available to support the development of both the channel and site water balances for the upstream drain. The channel water balance only was solved for the downstream drain to compare the different discharge responses of the leveed and non-leveed upstream and downstream drains.

The effects of the pumped groundwater into the drains (Section 4.5) have been excluded from the measured results as much as possible so these waters are not reflected in the water balances.

5.1 Drainage site

The groundwater catchment surrounding the upstream drain was defined as the area within which drain construction may lower the watertable. The distance within which this might have or could reasonably be expected to have occurred is the watertable zone of influence of the drain (Appendix B.1). Selecting a watertable zone of influence of 100 m each side of the drain produced a conceptual drainage site of 182 ha with a perimeter of 18 050 m (Fig. 26).

The extent of the watertable zone of influence was chosen from groundwater hydrograph analysis (Section 4.1). Using the relationship between watertable heads at increasing distance from the drain and groundwater inflows to estimate watertable zone of influence (Appendix B.3) did not produce the same corroborating results as for other drains (Cox & Tetlow in press).

Although the extent of the watertable zone of influence varies considerably, this is not important. This approach is to guide the choice of scale upon which to base the water balance in respect to drainage efficiency. If the drainage site area is too large the effect of the drain becomes insignificant and provides no comparable results with other drains. If the site is too small groundwater outflow into the drain may exceed the groundwater supply to the drainage site from recharge and aquifer discharge. This would cause an imbalance between the measured groundwater outflow (Q) and other variables of the water balance.

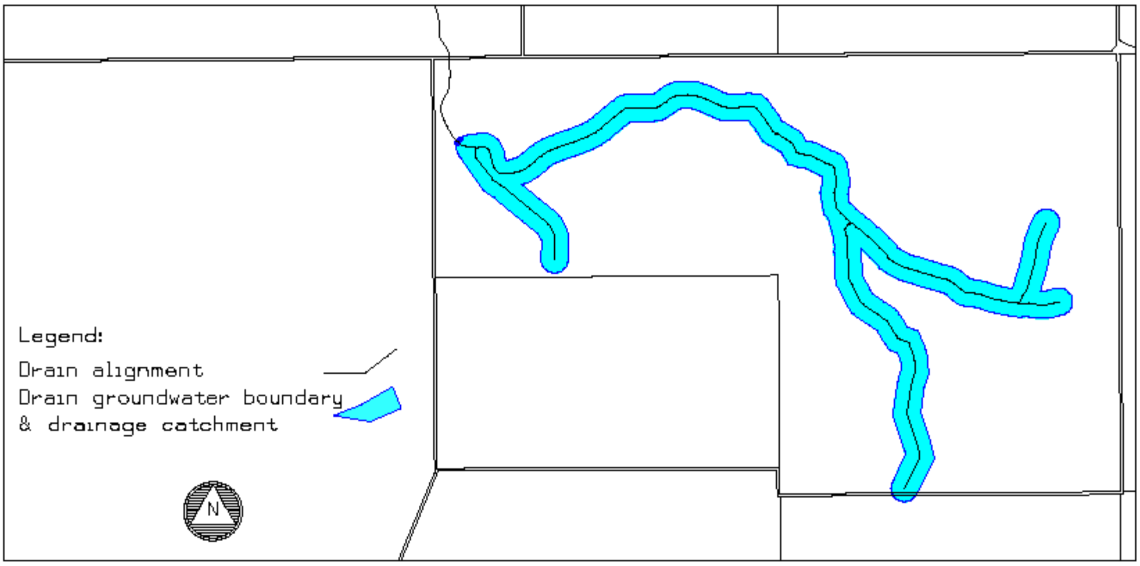

Figure 26 The upstream drainage site

5.2 Water balance for the upstream drainage site

A water balance for the upstream drainage site was produced from daily to monthly measured and extrapolated data (Appendix B) aggregated to produce a monthly time-step water balance. The comprehensive water balance for the 182 ha upstream drainage site is encapsulated by the equation:

$$P + A_D = (ET + RO + \Delta S_S) + \Delta A_S + G_E + G_O \qquad \textit{Equation 1}$$

Where: P: Rainfall (Section 2.1)

A_D: Aquifer discharge (Appendix B.6)

$(ET + RO + \Delta S_S)$: Rainfall losses by evapotranspiration, runoff and change in soil water (not measured)

A_S: Aquifer storage surrounding the drain (Appendix B.4)

G_E: Groundwater evaporative loss by capillary discharge (Appendix B.7)

G_O: Groundwater outflow from the site into the drain (Appendix B.2)

Substituting the measured and calculated values for the post-drain monitoring from 24 June 2004–31 December 2006 into the water balance equation gives (in kL):

$$1\,183\,000 + 115\,100 = 774\,300 + (-4000) + 376\,000 + 151\,700$$

Rainfall (P) provided the largest contribution of water (1 183 000 kL) to the drainage site and was mostly balanced by rainfall losses of evapotranspiration, runoff and the change in soil water storage ($ET + RO + \Delta S_S$). Rainfall losses (775 300 kL) accounted for about 65% of rainfall with groundwater outflow (G_O) into the drain 13% (151 700 kL). Rainfall losses in the equation were not measured values but were the product of the inputs to the water balance less the outputs, needed to make the equation balance.

The 376 000 kL of groundwater evaporative loss (G_E) is an expression of the potential loss of groundwater directly from the watertable by capillary rise and evaporation. This value is significantly overestimated because it is potential capillary rise, ignoring the limiting effects of evaporative loss or capillary discharge from the soil surface (Appendix B.7).

In practice, groundwater evaporative loss is a component of evapotranspiration and might normally be expressed as a rainfall loss component of the water balance, if at all. Groundwater evaporation has been included in this water balance as a mechanism to explain and quantify reductions in the watertable height beyond those that could be attributed to the drain.

Aquifer storage (A_S) reduced (negative value) by about 4000. It did not decline steadily but fluctuated through the monitoring period in response to groundwater moving in and out of the aquifer (Appendix B.4). Aquifer storage rose to its maximum in July 2004 of 7900 kL more than at the start of drainage. By April 2005 it had fallen to its minimum of 10 900 kL less than at the start of drainage. The results show aquifer storage change to be dynamic in response to the other water balance variables. Hence its stated volume was significantly affected by the timing of the start and end of the period of evaluation in comparison to changes in the other variables.

The water balance volumes are more easily compared when expressed as equivalent depths of water within the drainage site. The volumes from Equation 1 are expressed as their equivalent depth in millimetres with the largest value of 650 mm being for rainfall (Table 1). Rainfall (P) combines with aquifer discharge (A_D) of 63.3 mm to provide a total depth of 713.3 mm of water supply into the drainage site. The other four variables represent the loss of water from the site by evaporation, drainage and surface runoff. The loss of water by drainage is by groundwater outflow (G_O) of 83.3 mm from the drainage site into the drain (Table 1).

Table 1 Water balances expressed in mm depth within the drain site

	P	$\mathbf{A_D}$	**(ET+RO+ΔS_S)**	ΔA_S	$\mathbf{G_E}$	$\mathbf{G_O}$
24/6/04–31/12/06	650.0	63.3	425.5	-2.2	206.7	83.3
2005	249.1	27.1	203.4	-0.6	54.3	19.1
2006	294.0	19.3	197.0	0.2	76.3	39.8

Groundwater recharge was not shown in the comprehensive water balance although it was calculated. Recharge was embedded within the relationship between rainfall and discharge by groundwater evaporation and drainage. Amending the comprehensive water balance by removing rainfall and rainfall losses and directly incorporating recharge (U) results in the groundwater balance equation:

$$U + A_D = \Delta A_S + G_E + G_O \qquad \textit{Equation 2}$$

Aside from the addition of recharge (U) all of the variables shown for Equation 2 are the same as for Equation 1. The variables from the groundwater balance equation are conceptualised alongside the drainage channel in Figure 27.

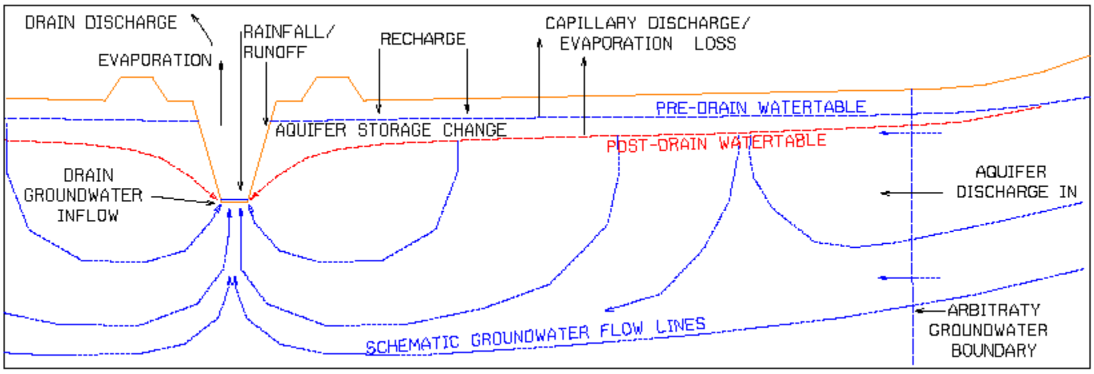

Figure 27 Conceptualisation of the water balances of the drain and surrounding site

The estimated groundwater recharge (Appendix B.5) for 2006 was more than twice that of 2005 (Table 2) in response to only 15% (44.9 mm) more rainfall in 2006 than in 2005 (Table 1). The more than expected recharge in 2006 was mostly in response to 139.2 mm of January–February rainfall (Fig. 3) causing an estimated 29.7 mm of the total 40.5 mm of recharge (Table 2).

Groundwater evaporation from Equation 1 was re-calculated in Equation 2 from the sum of the inputs ($U + A_D$) less the sum of the other outputs ($\Delta A_S + G_O$). This produced theoretical groundwater evaporation rates of about 25–50% of those from Equation 1. The reduced 2005–06 groundwater evaporation rates (Table 2) are more comparable with measured average annual upward groundwater fluxes of about 73 mm/yr (McIntyre 1982). These measured upward groundwater fluxes were in clay cores with watertables at around 0.5 m below ground level. The lower groundwater evaporation rates from the site compared with those measured could be explained by the slightly greater depths to the watertable beneath the site (Appendix A).

Table 2 Groundwater balances expressed in mm depth within the drainage site

	U	A_D	ΔA_S	G_E	G_O
24/6/04–31/12/06	70.6	63.3	–2.2	52.7	83.3
2005	17.8	27.1	–0.6	26.4	19.1
2006	40.5	19.3	0.2	19.8	39.8

5.3 Water balances for the drains

Groundwater outflow from the drainage site becomes groundwater inflow to the drainage channels (Section 5). The channel water balance for the upstream drain shows losses and accessions to inflow before its measured as discharge at the drain outlet:

$$Q_{US} = G_I + (P + RO) - \mathbf{E} \qquad \textit{Equation 3}$$

Where: Q_{US}: Measured discharge from the upstream drain

E: Evaporation from the drain channel

G_i: Groundwater inflow to the channel

(P + RO): Rainfall and runoff

These variables and interaction with the adjacent groundwater system are conceptualised in the drain and site water balances shown in Figure 27. The channel water balances were calculated daily with the results being aggregations of the daily values. Substituting the appropriate values into Equation 3 for the period 24 June 2004–31 December 2006 gives:

$$103\,800 = 151\,700 + 7200 - 55\,100$$

The discharge from the upstream into the downstream drain is shown in the water balance for the downstream drain:

$$Q_{TOT} = Q_{US} + G_I + (P + RO) - \mathbf{E} \qquad \textit{Equation 4}$$

Q_{TOT} in this water balance is the measured downstream drain discharge (Section 4.6):

$$306\,600 = 103\,800 + 221\,000 + 47\,700 - 65\,900$$

After subtracting the discharges from the upstream drain the water balance of the downstream drain can be expressed using Equation 3 as Q_{DS} being equal to:

$$204\,500 = 221\,000 + 47\,700 - 64\,200$$

Approximately 1700 kL of discharge from the upstream drain was lost between the upstream and downstream gauging stations. The daily water balances showed discharge from the upstream drain coinciding with no discharge from the downstream drain (Appendix CD 5.2b). The assumption is upstream discharge was lost to evaporation between the upstream and downstream gauging stations (Section 4.6).

The evaporation of the upstream discharge from the downstream channel creates a 1700 kL difference in evaporation from the total channel water balance (Q_{TOT}) compared to the downstream water balance (Q_{DS}). This also causes total discharge from the drainage scheme to be 1700 kL less than rather than the sum of the upstream and downstream discharges.

The total lengths of upstream and downstream drain channels are virtually the same (9250 m and 9420 m respectively) making their channel water balance results almost directly comparable. The most obvious difference is that downstream discharge is almost twice that from upstream. This greater discharge was driven mainly by groundwater inflow during 2005 and rainfall and runoff in 2006 (Table 3).

Table 3 Channel water balance results from Equation 3 in kL/km of channel

	Q	**E**	**G_I**	**(P + RO)**
Total Upstream	11 200	6000	16 400	800
Downstream	21 700	6800	23 400	5100
2005 Upstream	2400	1600	3800	200
Downstream	6000	3000	8600	300
2006 Upstream	5200	2900	7800	300
Downstream	8900	1900	7700	3100

The downstream water balance results are less robust than those from upstream. The upstream contributions to the downstream drain in combination with the 2600 m non-groundwater contributing section could have affected the interpolation of the other water balance variables that were derived from the drain discharges. The loss of about 1700 kL of upstream discharge from the downstream channel signals possible discharge losses from the other downstream drains. Hence, actual downstream groundwater inflows and evaporative losses could be higher than those reported.

5.4 Salt balance for the upstream drainage site

The mass of salt removed with groundwater outflow is the only value from the salt balance of the upstream drainage site that can be stated with any degree of certainty. The salt balance for the drainage site was estimated by calculating the mass of salt moved or removed by the groundwater variables for Equation 2. The groundwater variables of aquifer discharge, storage change and groundwater evaporation were assigned a salinity of 21 393 mg/L (Appendix B.2) to estimate their salt mass balances:

$$? + 2460 = (-90) + 2050 + 3270$$

Although the salt balance assumptions include that no salt is contributed to the site from other sources (rainfall and surface wash), the salt load from recharge is unknown, as indicated by the question mark (?). Salt is also incorrectly shown as being lost from the site by evaporation whereas actually it is not evaporated but transported to and deposited on the land surface and in topsoil by capillary rise and discharge. This accumulated surface salt can then be exported by surface runoff and/or re-infiltrated back to the watertable with the onset of recharge.

Taking these uncertainties into account there is a net reduction in salt load of the drainage site from 720–2770 t depending on the proportions of re-leached to exported salts. The net reduction is maximised if all of the 2050 t of salt transported to the land surface is exported by runoff together with the groundwater outflow by drainage: salt load from recharge is then zero. If all of the deposited salt is re-leached back to the watertable the recharge salt load equals groundwater evaporation, leaving a net salt export in groundwater outflow to the drain of 720 t.

The most important post-drain variables affecting the site salt balance are the inflow of salt by aquifer discharge and its outflow in drainage. The salts transported by the other variables of recharge (re-infiltration), capillary rise and discharge, and storage change appear either negligible by comparison or only represent the movements of salts between the other variables of the salt balance.

This salt balance is only indicative of the post-drain groundwater movements and associated mobilisation of salts. The salt loads of the groundwater variables could vary markedly from those estimated dependent on the comparative differences between actual and the assigned groundwater salinity of 22 393 mg/L.

For 2005–06 the average outflow of salt by drainage was equivalent to 6.35 t/ha/yr from the drainage site and the average inflow from aquifer discharge was 4.95 t/ha/yr (Table 4) confirming the net salt export. If the contributing average groundwater salinity from aquifer discharge were about 22% higher (27 500 mg/L) salinity inflows to and outflows from the site would be in equilibrium. With increasing aquifer discharge salinity the site salt balance would progressively show net salt accumulation as, was probably the case before drainage.

Table 4 Salt balances (t/ha) for the upstream drainage site

	U	$\mathbf{A_D}$	ΔA_S	$\mathbf{G_E}$	$\mathbf{G_O}$
24/6/04–31/12/06	?	13.53	–0.47	11.28	17.97
2005	?	5.79	–0.12	5.64	4.08
2006	?	4.12	0.05	4.23	8.63

5.5 Salt balances for the drains

The salt balance for the upstream drain reflects that the 3270 t of salts discharged from the drain outlet were sourced only from its groundwater inflows. By substituting salt loads into equation 3 assumes that the contributions of salts from rainfall to the channel are negligible and that any surface runoff of saline water into the channel is prevented by the drain levee banks (Section 4.5).

Upstream drain discharge and surface runoff provide additional sources of salt-flows into the downstream drain over and above that from its groundwater inflows. Surface runoff from the surrounding catchment into the downstream drain was in the order of 47 700 kL (Section 5.3) but its salinities were unknown and likely highly variable. Surface runoff mobilised salts from the surrounding catchment and into the drain thereby contributing to some peak salt loads associated with increased drain discharges (Section 4.6). Peak salt loads are also caused by salts that had accumulate within and were mobilised from the channels during runoff events (Section 4.6).

The mixing of salts from surface runoff with those accumulated and mobilised from within the channel reduced the ability to quantify the salt-flow into the drain from surface and groundwater inflows. Assuming a runoff salinity of zero the salt load discharged from the downstream drain (10 900 t) (Section 4.6) originated from groundwater inflows (7750 t) and discharge from the upstream drain (3270 t). The 120 t difference between salt inflows and discharges showed that some salt appeared to have been lost from or become entrained within the downstream channel (Appendix CD 5.2b).

Alternatively, 1860 t of salt was discharged from the downstream drain on days when surface runoff was thought to have occurred. By salt mass balance the average salinity of the 47 700 kL of surface runoff would need to be about 39 000 mg/L to have mobilise this amount of salt into the drain. If 1860 t of salt was transported into and discharged from the drain with surface runoff groundwater inflows were responsible for the remaining 5890 t. The range of salt-flow transported by groundwater inflow to the downstream drain is therefore somewhere between 5890–7750 t. Conversely, that transported into the drain by surface runoff is somewhere between 0–1860 t.

6 Discussion

6.1 Groundwater inflows

The random and single channel layout of the Pithara drains resulted in unbounded groundwater responses within the surrounding valley floor aquifer (Appendix B.1). Given the unbounded hydrology of the surrounding drained groundwaters it was not possible to differentiate the sources of groundwater inflows to the drains with any degree of certainty.

The pre-existing hydraulic gradients between the valley floor and hillsides suggested that groundwater would keep migrating towards the drain from the surrounding elevated lands within the catchment. The upwards heads of the deeper groundwaters beneath the valley floor and drain (Appendix A) provided corroborating evidence for this. The only restriction on the groundwater to the drain from aquifer discharge is as the transmissivity of the aquifer between the hillsides and drainage site.

Localised rain-fed recharge was conceptualised as intermittently mixing with aquifer discharge to provide a combined groundwater supply to the drain. The proportion of the groundwater supply that eventuates as inflow to the drain is affected by the characteristics of the surrounding aquifer and head differences between the drain channel floor and watertable alongside. Hence the drain is most efficient at removing groundwater when the surrounding watertables are highest. If the drain is to lower the watertable it must deplete its groundwater supply.

Drainage efficiency was largely impeded by the low permeability of the clay subsoils into which the drain channel was excavated. The water balances demonstrated that the upstream drain had about enough efficiency to remove groundwater equivalent to the volume of localised recharge or aquifer discharge, but not both. This left 'surplus' groundwater to continue to be lost by capillary rise and evaporation (Table 2) as probably occurred before drainage. For continued groundwater losses by capillary rise the watertable must stay close to the land surface.

Water quality analysis showed that the drain inflows were dominated by deeper groundwaters (Appendix B.3). Groundwater from hillside aquifer discharge is seen as the driver of deeper groundwater rise in the valley that provides sustained groundwater inflows to the drain during prolonged absence of rain-fed recharge. Although the semi-confined and unconfined aquifer characteristics are heterogeneous, groundwater flow towards the drain from various distances is seen as being radial rather than horizontal (Fig. 51). This could result in the flow paths of even localised recharge being radial via the deeper aquifer and not horizontal 'across' the watertable towards the drain. In support of this, sustained groundwater inflows mostly rose vertically through the channel floor and bottom 0.1 m of the batters and not through the sides of the channels.

Alternatively, there may be only limited mixing of groundwater from aquifer discharge and from in-situ recharge. The dense clay layers from about 1.0–2.0 m below land surface (Section 2.4) could provide some confinement between the shallow and deeper groundwaters. This might leave deeper groundwater rise to dominate inflows to the drain while in-situ recharge is confined to the shallow soil profile and lost mostly by capillary rise, evaporation and/or evapotranspiration. Only very close to the drain might the depressurisation of the deeper groundwater encourage recharge from the overlying watertable. This may explain the slight reduction in watertable salinities closer to the channel (Section 4.2).

The Pithara drain is functioning as a groundwater and salt outlet from its catchment where previously there was none. There is an irrefutable relationship between AMRR (accumulated monthly residual rainfall), the groundwater levels alongside and groundwater inflow rates to the drain (Appendix B.2). Under pre-drain conditions groundwater evaporated from large areas of the valley floor with salts accumulating in the topsoil and on the soil surface. With the onset of rain the salts were leached back into the soil or exported from the catchment by surface wash.

Salts now transported into the drains with groundwater inflows are directly exported from the catchment. Because the mass of salts exported in surface wash was not measured it is unknown whether the drain has overall increased salt export or has merely replaced or complemented surface wash as a means of export.

Regardless of the salt export processes on the valley floor the salt balance of the drainage site (Section 5.4) suggests that the drain will not effect change in long-term salt export rates. Aquifer discharge and upward groundwater rise ultimately regulate the mobilisation of salts to the valley floor. By reasoning, whether the salts are now exported by surface wash or drain discharge affects only the residence time and concentrations of the discharges, not the total salt loads.

6.2 Total drain discharges

The Pithara drains were built as leveed and non-leveed sections of approximately the same length to gauge differences in their performances. Groundwater discharges from both the drains were much modified by the effects of evaporation, rainfall and runoff from and to the channel. From the upstream drain average evaporative losses exceeded the rainfall and runoff contributions to the channel by about seven-fold (Table 3). For the downstream drain the contribution of about 45 000 kL of runoff from the surrounding catchment reduced the average proportion of evaporative loss to rainfall and runoff to about 1.3:1.

There was about 6000 mg/L difference in the total flow-weighted salinities between the upstream and downstream drains. The total flow-weighted salinities for the two and a half years of measurement were 31 500 mg/L upstream and 37 900 mg/L downstream. The higher average downstream salinity was mostly the result of greater evaporative losses from its non-groundwater-contributing creek section. This is confirmed by the 24 000 mg/L difference between the total and average daily flow-weighted salinities for the downstream drain compared to the 16 000 mg/L difference between the upstream values. The higher average daily salinities confirm that on most days the downstream drain discharges were far more saline than upstream ones, caused mainly by evapoconcentration of the drain flows.

There was no expectation that excluding or not excluding surface water runoff into the drain would have any effect on adjacent watertables. The rationale for excluding surface water runoff with levee banks was to reduce structural damage and channel sedimentation. The unregulated ingress of runoff into the deep drain conveys sediments, causes batter erosion and results in high discharges that scour and erode the channel (Cox et al. 2004). As a channel fills with sediment and becomes shallower its efficiency at draining groundwater decreases necessitating maintenance to restore its original depth.

Surface water runoff and rainfall into the downstream drain produced about 25% of its total discharge (Section 5.3) with more than half of this in January–February 2006. Less than 5% of the discharge from the upstream drain originated from runoff and rainfall. The 47 700 kL of runoff from 13 200 ha of catchment into the downstream drain was equivalent to an average runoff depth of about 0.4 mm or 0.06% of rainfall. During January–February 2006 the average depth of runoff was about 0.2 mm or 0.14% or the rainfall for the months.

The average annual depths of runoff were about half those of the greater Mortlock River catchment (1993–2002) of about 0.35 mm/yr (Mayer et al. 2005). The lower runoff from the project site can be explained by the high surface water losses within the surrounding catchment. The sandplain soils allow high infiltration rates on the valley flanks while runoff from the valley floor is detained within the many small playa lakes and saline depressions (Fig. 9). In all likelihood most of the runoff into the downstream drain rose from clayey and saline scalded areas fairly close (hundreds of metres) to the channel.

There was not enough runoff to expose any engineering advantages between the leveed and non-leveed drains as there was no noticeable difference in their sedimentation rates or other performance characteristics. Anecdotally, while runoff events are infrequent in the Wheatbelt, when they occur they are often intense and can cause great damage to the unprotected channels of non-leveed drains.

6.3 Groundwater response

In the months following construction the drain allowed some small reductions in the adjacent watertable that appeared to be related to the depressurisation of the deeper groundwater. The horizontal extent of the effect on the watertable was variable, within 50 m and sometimes to 175 m from the channel (Section 4.1). However, the initial reductions in height were often only centimetres in magnitude and did not appear enough to help in the reclamation of adjacent saline land.

During 2005-06 watertables within 400 m of the drain fluctuated in response to AMRR sometimes rising and remaining above their pre-drain levels. There were no consistent differences between the watertable fluxes within 100 m of the drain compared to those further away and in the comparison bores. No effective comparisons could be made between pre- and post-drain water levels because there were so few pre-drain measurements. A comparison of pre- and post-drain watertables might have shown a reduction in the equilibrium levels of the post-drain watertable in response to the depressurisation of, and so reduced upward leakage of, deeper groundwater.

Single unbounded drains cannot lower, let alone maintain, a watertable below a depth critical for farming, more so with the low soil permeability of the Pithara site (Appendix B.1). Being unbounded, the Pithara drains have become part of the catchment valley floor groundwater drainage system and act in conjunction with capillary discharge to maintain watertables below the land surface. From the water balance, 83.3 mm was drained and 52.7 mm was evaporated from the drainage site (Table 2).

Based on these figures the watertable would theoretically need to be at 1.05 m below ground level to enable the capillary rise and discharge of the total 136 mm (83.3 + 52.7). This 1.05 m average depth to watertable was derived using the process for estimating capillary rise, but in reverse (Appendix B.7). If compared with the measured average depth to watertable (1.36 m) across the drainage site, the analysis suggests an average 0.31 m lower watertable as a result of the drain.

Using the same analysis, the 2005 measured average watertable level of 1.51 m below ground might have been 0.32 m higher without the 19.1 mm drainage from the drainage site. For 2006 the capillary discharge of the 39.8 mm drained equated to a watertable 0.34 m higher than the 1.33 m drained average depth below ground level for that year.

The potential drain effects on the watertable cannot be adequately estimated from changes in the rate of capillary rise but show the potential drain contribution to a lower watertable. The closeness of these estimated watertable reductions attributed to the drain are coincidental or the result of spurious data. There should be no relationship between the drainage rate and rate of capillary rise even though both processes are at least partially controlled by the watertable level. The height of the watertable above the drain base affects the drainage rate (Appendix B.3) and the depth of the watertable below ground affects capillary rise (Appendix B.7).

The loss of groundwater from capillary discharge of the capillary rise is also largely controlled by the potential evaporation rate. Consequently, the timing of upward fluctuations in the watertable with regard to the timing of high or lower evaporative potential has a large influence on the loss of water by capillary discharge (Appendix B.7). Hence, differences in evaporative potential affect the consistency in the relationship between capillary discharge, drainage rate and watertable levels.

With watertable responses to both drainage and capillary rise intrinsically but not directly linked it becomes nearly impossible to attribute watertable responses to one or the other. It could be argued that, in the absence of drainage, watertables and so capillary discharge rates would be higher so watertables would fall faster in response to capillary discharge alone. Conversely, without capillary discharge a greater proportion of the groundwater could be drained and so watertables would fall in response to only the drain.

It could be postulated that as long as the post-drain watertable remains close to the land surface the continuing potential for capillary discharge will conceal the effect of the drain on the watertable level. A watertable remaining near the land surface alongside the drain shows that the groundwater supply exceeds the drainage rate because groundwater continues to be lost by capillary discharge. The watertable level alongside the unbounded drain is unlikely to fall because the supply of groundwater is not 'designed' as for bounded drains (Appendix B.1) and so will always exceed the drainage rate. Under these conditions the 'surplus' groundwater will need to continue to be lost by capillary discharge. For capillary discharge to continue the watertables alongside the drain must remain elevated. The apparent lack of significant watertable reductions at Pithara is thought to conceal that the post-drain watertable is controlled by both drainage and capillary discharge rather than by capillary discharge alone.

6.4 Salt land recovery

Capillary rise and discharge are the processes by which solutes are transported to and accumulated within the root zone and land surface at Pithara. When the groundwater rises to within about 1 m of the land surface the potential rate of discharge increases more than proportionally in response to further watertable rises (Fig. 53). Hence the closer the watertable is to the surface the greater the potential for land salinisation. It is when watertables are nearest the land surface that any lowering effect by the drain has the greatest influence on reducing capillary discharge and its associated upward salt fluxes.

In the period June 2004–December 2006 using a salt balance for the drainage site the average topsoil deposition from the upward flux of salts was equivalent to 4.51 t/ha/yr (Table 4). There was no estimate of the upward salt fluxes under pre-drain conditions due to the lack of sufficient pre-drain watertable measurements. The salt balance provided no indication of the overall change in soil salt storage levels caused by the drain because only the salt export by the drain was measured.

Any fall in the watertable will reduce the upward flux of salts by capillary discharge. Topsoil salt accumulation will fall if the upward flux of salts becomes less than the downward flux in response to leaching and/or the loss of salts by surface runoff.

Significant and consistent salinity reductions in the upper 35 cm of the soil profile were measured at transect 4 from June 2004–April 2009 (Bell et al. 2009). Soil salinities at 5, 15, 25 and 35 cm depth were measured as Ec_W 1:5 soil water extracts. Average soil salinity was about 175 mS/m with a range 125–240 mS/m for all of the samples taken in June 2004. By

October 2006 average salinity had fallen to about 140 mS/m with individual samples 80–240 mS/m. The reduction was most noticeable in the upper 15 cm of the soil profiles with almost no reduction at 35 cm. April 2009 salinities averaged 100 mS/m with levels around 75 mS/m in the upper profile increasing to 175 mS/m at 35 cm. The soil salinity decline from 2006–09 was more consistent throughout the full 35 cm depth of the profile.

Average salinities at a nearby control site increased from about 175 to 225 and then 260 mS/m through the 2004, 2006 and 2009 sequence of measurements. Although starting at similar levels in 2004 it appears that drained soil salinities had reduced by about 75 mS/m while those undrained had increased by 50 mS/m. Notably, most of the salinity changes in the drained and undrained conditions occurred within the upper part of the soil profiles.

There is potential for high variability in soil salinity measurements related to for example different sampling times during the year and in relation to preceding rainfall (Bell et al. 2009). While the results show that salinity levels are significantly reduced at the transect 4 bores, ongoing measurements or other indicators of reduced soil salinity are needed to corroborate these results.

The soil salinity reductions have changed the status of the sample site from ‘extremely saline’ to ‘highly saline’ (Moore 1998). A barley crop established at the site produced suboptimal emergence in June 2006. Barley crop emergence increased with proximity to the drain, with maximum emergence at 25 m from the end of the plots closest to the drain (Fig. 28). The emergence rates at 25 m were 40% below the target plant density for good yields while emergence at 50 and 75 m from the plot end were well below optimum for barley crops (Bell et al. 2009).

Cereal crops only return satisfactory yields on highly saline land when seasonal conditions are most favourable (Moore 2008). The post-drain change in land salinity status may provide the opportunity to introduce some salt-tolerant clover and medics in combination with barley grass rather than the barley grass, halophytes and bare ground that persisted before drainage.

Ongoing reductions in topsoil salinity as occurred from 2006 to 2009 may have led to further improvements in potential crop and pasture productivity, particularly under favourable seasonal conditions. This could be determined at sometime in the future by re-sowing barley at the trial site.

Figure 28 Sub-optimal 2006 barley crop at transect 4

6.5 Drain performance simulation

Steady-state drainage comparisons

The Hoodghoudt steady-state drainage equation (Ritzema 1994) was used to investigate ***potential*** watertable drawdown responses to the upstream Pithara drain. This equation is commonly used for drainage design to predict watertable positions and groundwater inflows in response to various recharge conditions. When used as a drainage planning tool the results from the equation guide the design of the appropriate spacing between parallel drains.

The equation's main input value is a nominated maximum watertable height midway between parallel drains. The nominated watertable height is a function of the proposed land use or required cropping conditions. The model calculates the appropriate drain spacing required to maintain the watertable at the nominated height (Section 5.1) by combining this value with those of the aquifer parameters, drain depth and recharge. The highest point in the watertable between the drains approximates reality so is always assumed to remain at the midpoint between them (Fig. 40). Hence the model solves for the distance between the drains in relation to the maximum watertable height midway between them.

The midpoint of the watertable between parallel drains has been equated with the outer limit of the watertable zone of influence of the unbounded drain (Fig. 41). By contrast with parallel drains the comparable extent of the watertable zone of influence is not easily defined or predicted for single unbounded drains (Section 4.1). The distance and height at which the drawdown curve ends and is replaced by the natural gradient of the watertable was not easily identifiable from the drawdown curves (Appendix A).

The steady-state equation is not normally used for drainage design or comparison where the variables of both watertable height and extent of the watertable zone of influence are dynamic or unknown. To overcome this for Pithara the deeper groundwater heads were substituted for the 'mid-point' watertable height in the equation. This was considered an acceptable approach given that much of the drain discharge originated from the head-driven upward movement of the deeper groundwater (Section 6.1). This substitution of groundwater heads for watertable heights enabled the model to be used to explore the interactions between recharge, watertable heads and zone of influence, and drain discharges for the unbounded drain.

The model assumes the presence of homogenous aquifer conditions which is contrary to the heterogeneity at Pithara as demonstrated by the range in aquifer hydraulic conductivities, thicknesses and groundwater responses and qualities. For modelling purpose averages of these values were used to simulate homogeneous conditions. Average hydraulic conductivity of 0.036 m/d and aquifer thickness of 15 m were combined with average groundwater heads that existed for the respective periods of simulation in Table 5.

Based on average watertable heads and groundwater inflows (Appendix B.3) from the whole period of monitoring the simulated watertable zone of influence was 96 m with a drainage rate equivalent to 34 mm/yr (Table 5). In Table 2 the groundwater outflow is the equivalent of a drainage rate of about 33 mm/yr for the same period. The drainage site was calculated from within 100 m each side of the drain (Section 5.1) which closely corresponds with both the 96 m simulated and estimated watertable zones of influence (Section 4.1).

Table 5 Steady-state drawdown distance and drainage rate calculated from average groundwater heads and inflows

Simulation period	Groundwater head (m above db)	Groundwater inflow (kL/d/km)	Watertable ZOI (m one side)	Drainage rate (mm/yr)
Average total	1.20	18.0	96	34
2005	1.07	10.4	123	15
2006	1.23	21.4	88	45
Min recharge (Nov-05)	1.17	5.9	177	6
Max recharge (Jan-06)	1.21	32.2	68	87
Min head (Dec-06)	0.87	4.8	162	5
Max head (Aug-04)	1.56	53.6	59	157

As heads fell nearer to the drain base level the theoretical watertable zone of influence widened as groundwater was drawn from increasing distances from the drain. The line of best fit that represents the relationship between head and zone of influence begins to flatten noticeably when heads are about 1.4 m above the drain base and the watertable zone of influence is at 60 m (Fig. 29). The flattening of the line beyond this point shows a propensity towards increasing horizontal extent of the zone of influence in favour of further reductions in watertable heads.

Groundwater inflow to the drain of about 45 kL/d/km corresponded with a watertable zone of influence of 60 m. Under these conditions the volume of water drained from within the watertable zone of influence was equivalent to a drainage rate of 137 mm/yr. A 0.3 m reduction in groundwater head resulted in doubling the watertable zone of influence and reducing groundwater inflow to about 11 kL/d/km. The resultant drainage rate was equivalent to about 17 mm/yr.

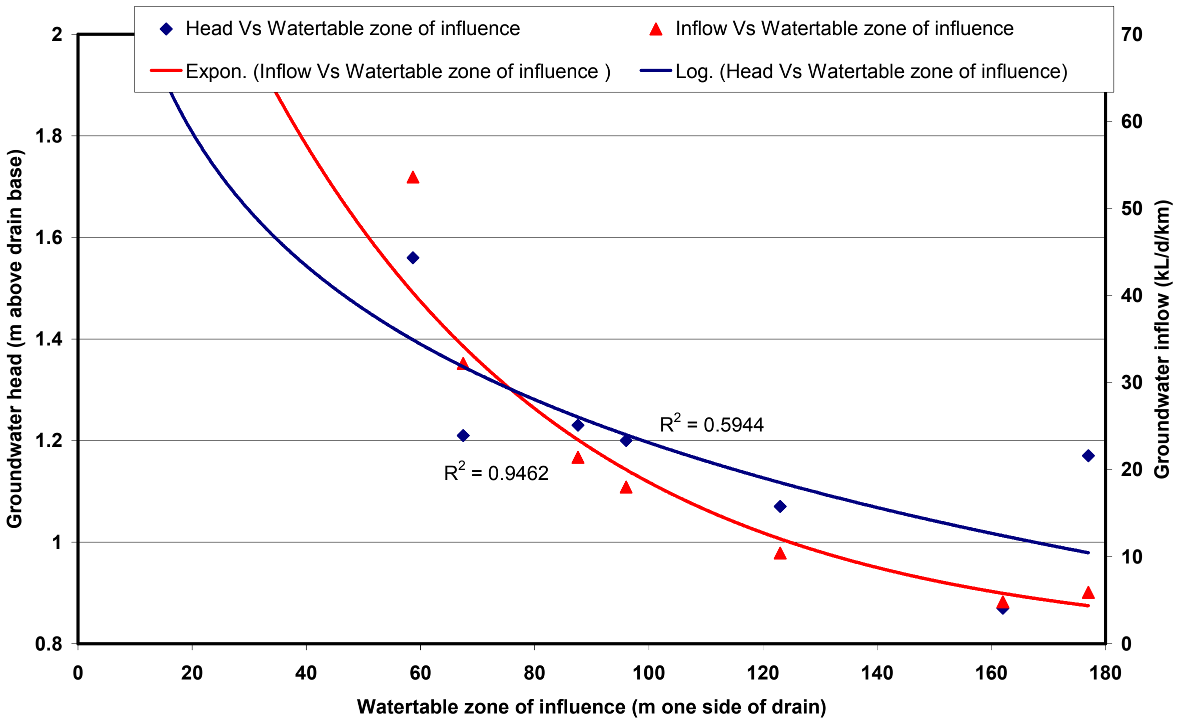

Figure 29 Relationship between head, groundwater inflow and watertable zone of influence from Table 6

A potential watertable zone of influence of about 60 m each side of the channel with groundwater heads at about 1.4 m above the drain base should be easily identifiable from groundwater hydrographs. Further increases in the watertable zone of influence would become less obvious, being associated with smaller reductions in the watertable height.

The groundwater inflows used in the simulations were derived from the water balances (Section 5.2) for the various periods of analysis to provide an indication of real drainage capability. Because the groundwater inflows to the drain were used instead of groundwater supply to the drainage site there may be large discrepancies between the simulated potential and observed watertable zones of influences. A reducing watertable zone of influence coincides with an increasing drainage rate (Table 6). If the drain were required to drain the entire volume of groundwater supply from the drainage site most of the zones of influence quoted would be reduced by about 50% in response to the increased in drainage rates. The average watertable zone of influence reduces down to 45 m from its previous 96 m extent.

These differences in watertable zone of influence highlight the deficiencies of the unbounded approach to drainage in controlling watertables. While the drain was evaluated within the context of its actual drainage capacity it could develop a watertable zone of influence in the order of 100 m each side. Once subjected to the full supply of available groundwater the watertable zone of benefit contracts in response to the rise in watertable caused by the excess supply.

Making drains work

The simulations highlight the inadequate performance of unbounded drains which stems from the inability to control the supply of groundwater to their drainage catchments. By constructing parallel (bounded) drains the watertable zone of influence can be restricted to achieve a balance between groundwater supply to the drain and drainage efficiency. In doing so aquifer discharge is largely disconnected from the groundwater system between parallel drains leaving only the drainage of in-situ rain-fed recharge to achieve watertable control. Under these conditions it is possible to induce more vertical than horizontal changes in the watertable position.

Nulsen (1981) suggests that watertable rises higher than 1.5 m below ground level cause yield reductions in barley crops as a result of dryland salinity. Adopting 1.5 m as a maximum watertable height in conjunction with the aquifer characteristics from above enabled drain spacing for adequate watertable control at Pithara to be calculated.

In the absence of aquifer discharge the rain-fed recharge values can be input directly into the steady-state equation as drainage coefficients. With a drainage coefficient equal to recharge of 28.2 mm/yr (Table 2) parallel drains spaced 188 m apart will maintain the watertable at a minimum depth of 1.5 m below ground level. To control watertable rises only in response to the lower 2005 recharge of 17.8 mm drain spacing could be increased to nearly 250 m, whereas for the higher 2006 recharge of 40.5 mm drain spacing would need to be reduced to about 150 m.

A drain spacing of about 160 m would be needed to control watertable rise in response to the 37.1 mm of recharge generated from the average 1970–2007 rainfall of 353.1 mm (Section 2.1). The 37.1 mm recharge rate is equivalent to 10.5% of rainfall (Appendix B.5). By designing drain spacing based on average rainfall and recharge theoretically results in watertables rising above the 1.5 m target depth during 50% of years. To completely prevent the risk of watertable rise above the target depth a drain spacing of 110 m is needed to control expected recharge of up to 70.6 mm/yr. This recharge might have occurred in response to the maximum annual rainfall in 1999 of 672 .4 mm (Fig. 2).

Parallel drain design can include a risk assessment of the probability of recharge exceeding drainage capacity and the impacts in terms of crop losses. At Pithara, parallel drains spaced at 140 m could reliably control the watertable below 1.5 m for 90% of years. This result is based on a drainage coefficient of 47.7 mm/yr calculated from 1970–2007 annual rainfall and expected recharge. Annual recharge from only four out of the approximate 40 years of rainfall record were calculated to exceed this drainage coefficient and so 90% level of watertable control reliability.

Because recharge is not uniform within the dryland cropping environment the risk assessment approach results in relationships between risk, recharge and drain spacing. Although the examples provided above are based on annual recharges, groundwaters have been shown to be most responsive to monthly rainfall and recharge (Appendix A). On an annualised basis monthly recharge ranged from an equivalent 0–297.6 mm/yr during the life of the project (Appendix B.5).

The use of drainage coefficients based on these more intense monthly or daily recharge values could lead to the over design of Wheatbelt drainage with the steady-state equation. Factors such as soil and aquifer storage play an important role in attenuating watertable rises in response to recharge. This might mean that recharge from intense events may not need to be drained instantaneously, but rather can be temporarily stored within the aquifer and drained over time. Other drainage models such as the unsteady state equation (Ritzema 1994) are better able to simulate the affects of recharge on changes in aquifer storage and drainage to better predict watertable positions.

Soil permeability is a major contributor to drain spacing with soils of higher hydraulic conductivity contributing to wider drain spacing. Given a hydraulic conductivity of 0.25 m/d rather than 0.036 m/d the drain spacing in response to average recharge could be increased from 188 m to about 550 m. A scheme with drains at wider spacing will cost less per unit area drained and increase the likelihood of the scheme being cost effective.

6.6 Construction costs

The total construction cost of 13 960 m of 2.5 m deep drainage (Section 2.5) and associated works was $116 372 (ex 10% goods and services tax). The sum is the cost of the drainage scheme as it would have been constructed by the landowner and excludes monitoring and additional costs associated with the Engineering Evaluation Initiative.

Excavation of the drain channel and the construction of the levee banks was the largest construction cost item at $90 079, an average cost of $6453 /km of drain. Culvert pipes for the 11 crossings (Appendix CD 3.1) for the drain cost $18 534 plus $1473 delivery. Culvert installation costs within the Petrador Farms property are included in the above drain construction costs. Installation of the culvert under Pithara East Road was completed with supervision, traffic control and assistance from the Shire of Dalwallinu at a subsidised cost by the Shire of $3522.

7 Conclusions

The large single drainage scheme at Pithara led to some reductions in the watertable through the drainage of groundwater. Before drainage lowering of the watertable was caused by the evaporation of groundwater from the valley floor, resulting in salinisation. After drainage, the watertable was lowered by the combined effects of both groundwater drainage and evaporation.

Reductions in topsoil salinity and waterlogging with associated improvements in the productivity of the land are expected to result from lowering the watertable by drainage. However, recovery of salt-affected land sufficient for dryland cereal cropping is unlikely. The watertable reductions were minor and inconsistent, and easily counteracted by subsequent recharge events. Topsoil salinity will not decline substantially and conditions will remain suitable for only salt-tolerant crops and pastures.

The inability of the Pithara drain to substantially lower the watertable is partially a consequence of its inadequate drainage capacity compared with the vast body of groundwater within the catchment. Although the drain has provided a ‘flowing’ groundwater outlet for the catchment the volume discharged is inconsequential to catchment wide recharge and aquifer storage.

The proven method of achieving watertable reductions with drainage is to design schemes that compartmentalise the groundwater system between drains. With parallel drainage schemes recharge between the drains is balanced against drainage capacity designed to lower watertables. Regrettably, achieving adequate watertable control at Pithara might demand a drainage density that would interfere with the broad-acre farming practices. The main limitation parallel drainage in this landscape being the low permeability of the aquifer and overlying soils.

Appendix A Watertable responses

Appendix A.1 Comparison bores

Of bores 024–027 originally selected as comparison bores, 026 and particularly 027 seem to provide the best comparative undrained groundwater conditions corresponding with the drainage site. Placed at the transition between slight–moderately to severely salt-affected land meant the watertable within these bores fluctuated through a range of 0.3–1.5 m below ground level (Fig. 30). This range of fluctuations was similar to that measured in many of the drain monitoring bores.

Shallow groundwater level fluctuations are consistent with trends in the normalised AMRR. After May 2004, groundwater levels are seen to rise rapidly and after May 2005, gradually in response to the onset of winter rainfall. Subsequent recessions in the groundwater hydrographs started around September each year, corresponding with declining winter rainfall and increasing groundwater evaporation (Fig. 3).

Groundwater levels in all the bores again rise rapidly by 0.5 m from December 2005–January 2006, consistent with a 70 mm increase in AMRR. These rises were largely driven by about 50 mm of out of growing season rainfall during mid January 2006 (Fig. 3). This was followed by a delayed onset of the normal rainfall pattern, allowing groundwater levels to recede before the onset of winter rain in August.

Although groundwater level trends are consistent with AMRR, the magnitude of groundwater rises are affected by the timing, intensity and duration of rainfall events. For example, even though the 2005 annual rainfall was 47 mm greater than for 2006, watertables rose higher in 2006 due to the preceding high January 2006 rainfall. Upward watertable fluxes such as this could have a profound effect on topsoil salinities, drainage efficiency and the subsequent ability to recover saline land.

Bores 026–027 produced fairly uniform and ‘as expected’ responses to climatic changes. The last groundwater levels measured in December 2006 were at or slightly above those measured at the time of drain construction before the onset of winter rainfall in June and July 2004 (Fig. 30).

Although it did not appear affected, 024 was not used as a comparison bore due to eventual close proximity of the newly constructed drain (Fig. 10). Bore 025 was also not used because fluctuations in water levels were possibly affected by enhanced recharge from leakage from an adjacent shallow drain. The 0.5 m deep drain channel was within 10 m of 025, and was often remained filled with water for several weeks after prolonged rainfall.

The deeper groundwater heads below comparison bores 026–027 were not measured due to the absence of paired deep bores. Since the shallow comparison bores are within discharge environments there is the possibility that the watertable level is influenced by upward leakage from the deeper aquifer. The only measurement of deeper the groundwater level was taken from a 37 m deep production bore (037) located midway between the comparison bores

(Appendix CD 3.2). At the initiation stage of the project in December 2003, the water level in 037 was about 0.6 m below ground level.

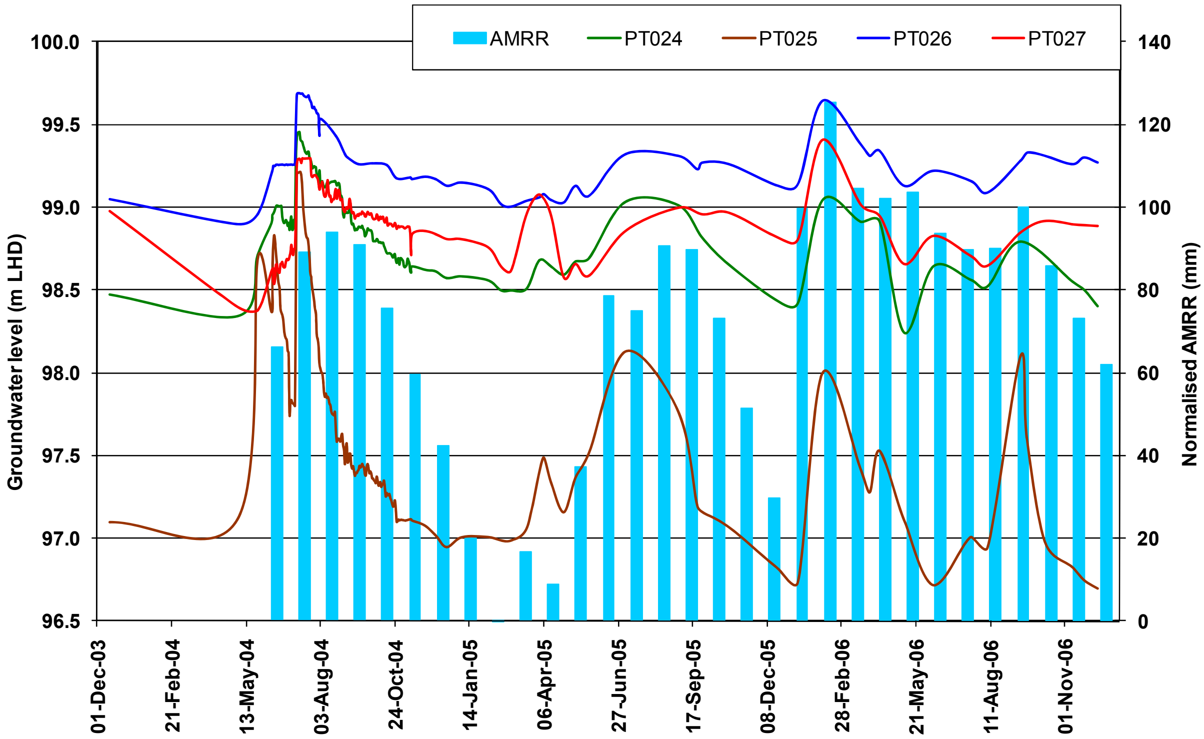

Figure 30 Comparison bore hydrographs with normalised AMRR for the site

Appendix A.2 Transect 1

Bore transect 1 (001–006) is perpendicular to and midway along a 1 km tributary drain. (Fig. 10). The bores are 20 m (001 and 001D) to 400 m (006) from the drain. The deep bore in this transect intersected basement at about 15 m below ground level (Appendix CD 2.5).

Transect 1 is surrounded by flat severely salt affected land valley with many small depressions that inundate during winter. The topsoil is a heavy grey clay (silty) which tends to disperse with rainfall. An open pit dug close to the drain alignment contained no groundwater on the first day but from the subsequent very slow inflow was estimated to have a hydraulic conductivity of 0.007 m/d. Slug withdraw tests on bore 002 produced slightly better conductivity results of 0.038 m/d.

Regrettably, drain construction at this transect coincided with a period of significant natural groundwater fluctuations that tended to obscure watertable reductions that might have been caused by the drain. Only in 001 was there a watertable response immediately after drain construction – a just detectable steepening in the downward trend in the hydrograph during the two weeks after drain construction (Fig. 31).

The transect 1 hydrographs showed a downward trend throughout the post-drain monitoring period: slightly greater for the closest bores and least for the bore 400 m from the drain. Hydrographs for bores in between appear to decline at about the same rate. Linear trend lines drawn through the hydrographs showed that groundwater in bore 001 at 20 m from the drain declined by 0.38 m while in bore 006 at 400 m declined 0.09 m. If bore 006 is representative of a comparison bore, the overall groundwater decline over the post-drain monitoring period for bore 001 is 0.29 m (0.38–0.09 m).

During periods of prolonged groundwater regression the hydrographs in the bores at up to 100 m show a slightly increased rate compared to those further away. At 275 m and beyond, there is no discernible difference in the hydrographs of the transect bores and of the comparison bore 027. The drain appears to have a small influence on groundwaters within 100 m – less than the seasonal trend. There is a general divergence of bore water levels during the summer and a convergence in winter and after heavy or prolonged rain.

The deep groundwater level (001d) trend coincided with that of the watertable level (001). The few pre-drain measurements showed the water level in the deep bore was about 0.2 m above the watertable (Fig. 31). This separation between deep and shallow groundwater levels continued post-drain but with some seasonal influences. In general, during summer when evaporation is dominant, the groundwater level is higher than the watertable. In contrast, during the wet winter of 2005, the watertable rose above the groundwater level. Between April and August 2005 the site had double the rainfall of the same period in 2006. Both before and after installation of the drain this site continued to be a groundwater discharge site except during periods of heavier rainfall and recharge.

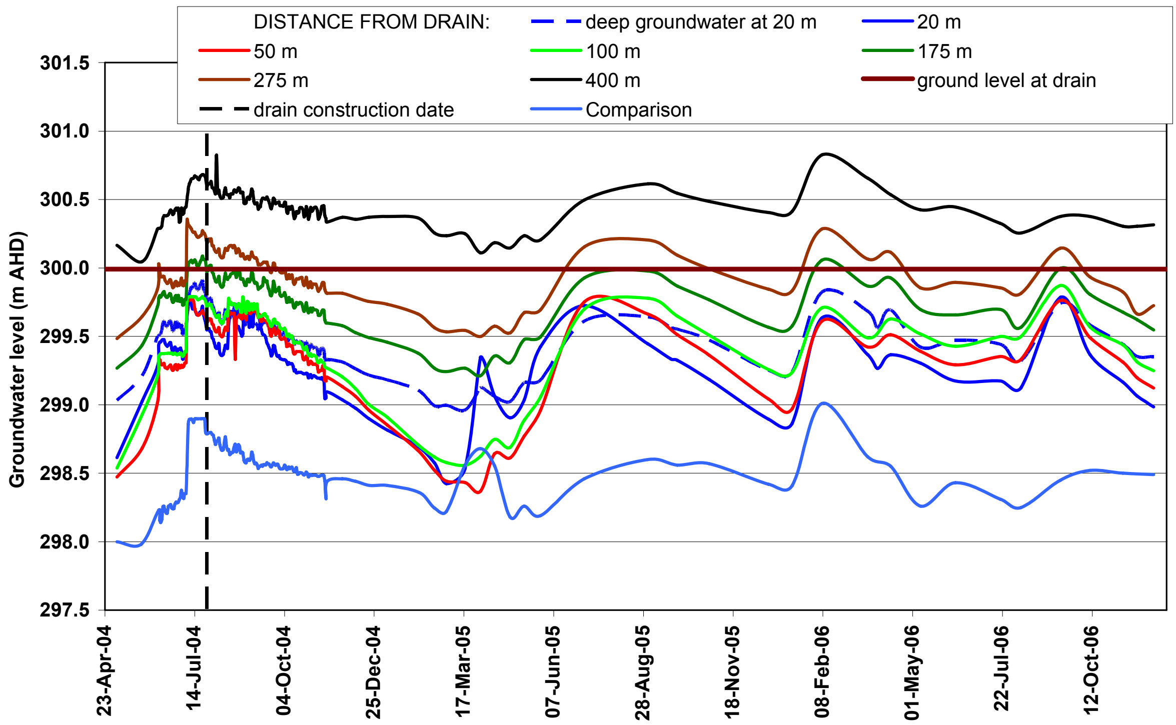

Figure 31 Hydrographs of transect 1 bores (PT001–PT006) and comparison bore 027

Four watertable profiles, on 16 June 2004, 18 March 2005, 7 February 2006 and 7 December 2006 respectively, are pre-drain (June 2004), subsequent post-drain lowest and highest and last measured watertables (Fig. 32). The 16 June 2004 watertable was the lowest pre-drain watertable and approximated the surface topography. The lowest post-drain watertable occurred at the end of a dry summer.

The inconsistently high watertable in bore 001 on 16 June 2004 and 18 March 2005 was most likely a response to recharge from ponded water in the drainage depression and alongside the levee, following rainfall. The localised difference in watertable heights reflects the low hydraulic conductivity at this transect. The highest watertable level was in response to the greater than 50 mm of rainfall during January 2006. By December 2006 the high watertable has fallen back to below its pre-drain level.

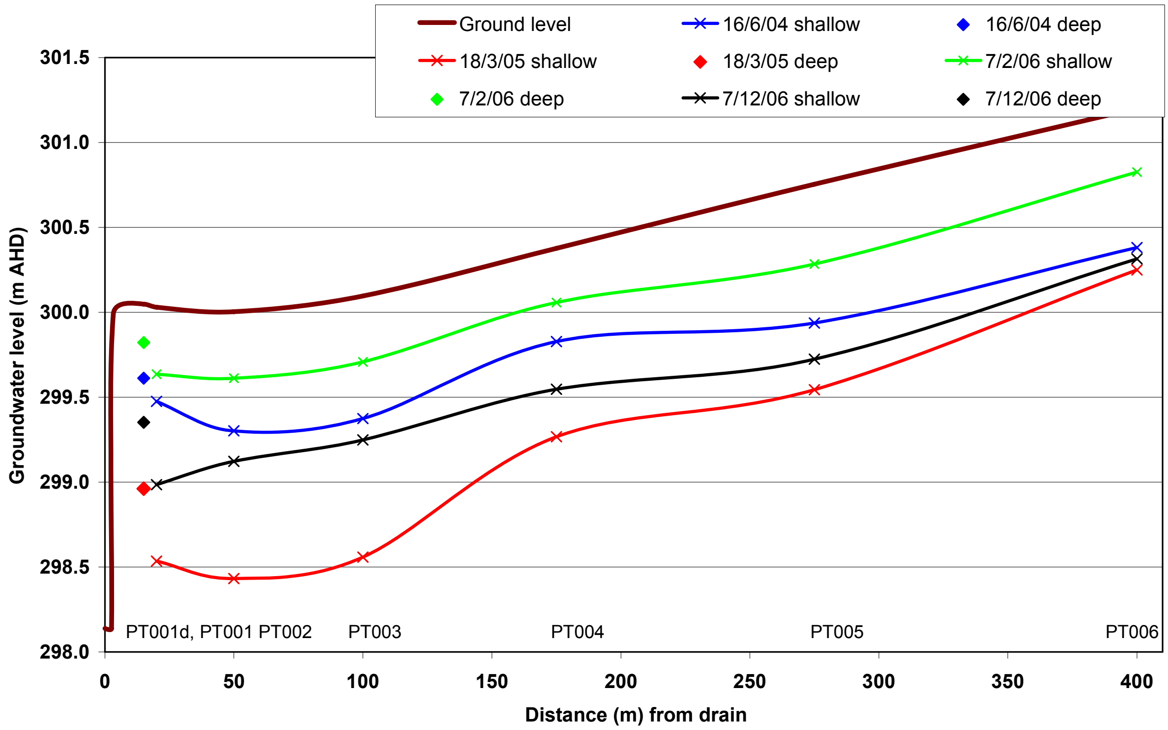

Figure 32 Transect 1 watertable profiles

Appendix A.3 Transect 2

Bore transect 2 (007–011) is perpendicular to the main drain and 1.2 km upstream of the upper gauging station (Fig. 10). The valley floor narrows at this transect (Fig. 9) and only bores at 20 and 50 m from the drain were surrounded by flat salt-affected land (Fig. 5). Bores from 100–275 m (009–011) are on a sandplain hillside with gradient of about 1%. The furthest bore at 275 m is 2 m higher in elevation than the bore at 20 m.

The hydraulic conductivity measured from bore 008 (50 m from the drain) was 0.308 m/d. This is an order of magnitude higher than the other measured bores and the 0.05 m/d estimated hydraulic conductivity from inflow into an open pit dug at the drain alignment. Anecdotal evidence suggests the reason for this higher hydraulic conductivity is that bore

008 is in more permeable soils along the footslope or edge of the flat valley floor (Section 2.5).

As it was excavated past transect 2 the effect of the drain was to reduce the rate of groundwater rise or increase its rate of decline at up to 175 m. The drain had an immediate effect on lowering both the watertable and groundwater level at 20 m on 22 June 2004 (Fig. 33). At 50–100 m the drawdown is almost imperceptible, but there is a noticeable flattening of the pre-drain rising hydrograph at 175 m, after drain construction. The comparison bore hydrograph showed a fairly uniform rising trend until the onset of recharge on the 11 June 2004.

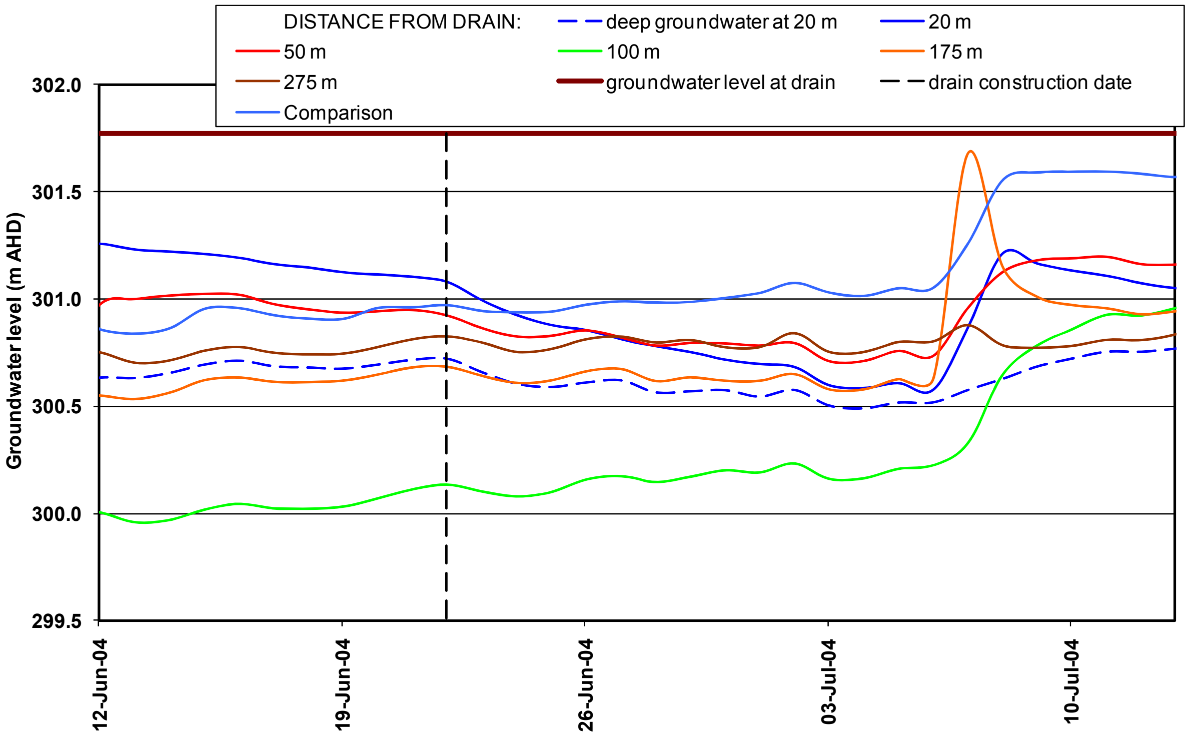

Figure 33 Hydrographs of transect 2 bores at drain construction (007–011) and comparison bore 027

On 11 June 2004 11 mm of rainfall caused the water levels in the valley floor bores at 20 and 50 m to rise by 0.3 m while the water levels in the other bores on the hillside were not affected (Fig. 34). This localised groundwater rise was believed to be caused by leakage from a shallow drain (Section 3.1) constructed adjacent to the groundwater drain. The enhanced recharge from the leakage caused the watertable within 50 m of the drain to rise above that of the remaining transect (Fig. 35). Within a month (10 July 2004) of the recharge the watertable along the transect had both risen and flattened, but still sloped slightly away from the drain.

Leakage from the shallow drain was noticed to have again caused accelerated watertable rise close to the drain with the onset of winter rainfall and runoff in June 2005. This is shown as the watertable hydrographs for bores at 20–50 m drain rising above the hydrographs of the outer transect bores (Fig. 34).

Linear trend lines imposed over the post-drain hydrographs showed a downward trend for all bores. This may be naturally skewed by the high rainfall at drain construction and below-average rainfall towards the end of the monitoring period (Section 2.1). The downward trend was greatest in the bores at 20 m and 50 m while the declines were much less in bores further from the drain.

In general, the fluctuations in all bores were similar. Compared to the comparison bores, levels fell by approximately 0.25 m more (by 1 m) from July 2004 to May 2005 in response to evaporation, drainage and/or natural drainage. In May 2005, they began to rise steadily with the onset of winter rains until 140 mm of rain fell in January and February 2006, causing a sudden rise of about 0.5 m (Fig. 34). Despite high evaporation rates at the time the high water levels took three months to fall to close to their pre-summer levels. By May 2006, bore levels at 175 and 275 m were still 0.3 m higher than pre-summer. From June 2006, levels in all bores declined slowly so that by December 2006 they were the same as in December 2005.

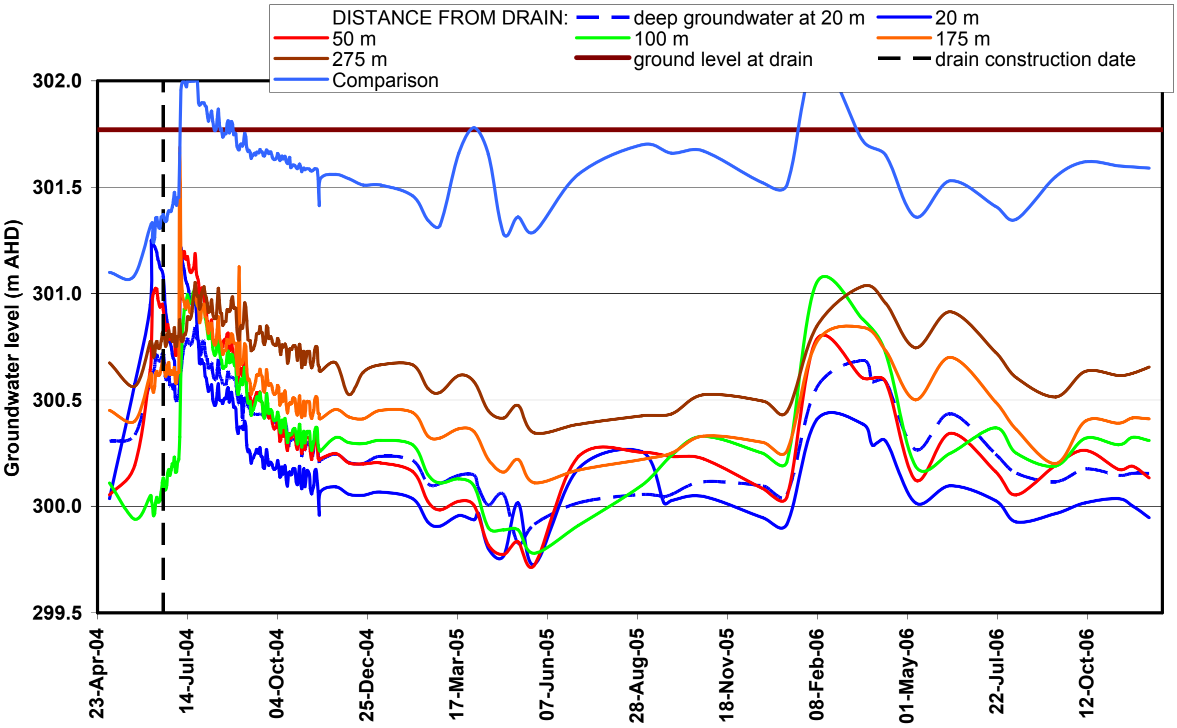

Figure 34 Hydrographs of transect 2 bores (007–011) and comparison bore 027

The deeper groundwater trend (007d) generally mirrored that of the watertable. The head within the deep aquifer (007d) remained approximately 0.2 m above its pair 007 watertable most of the time (Fig 34) except during winter in 2004 and 2005. It appeared that the watertable (007) could rise above the groundwater level (007d) in response to localised recharge caused by leakage from the shallow drain. Where recharge was more widespread, the groundwater head remained above the watertable as in February 2006 (Fig. 34). The watertable stayed below the groundwater level in winter 2006, probably as a result of the dry winter and the absence of localised recharge from the shallow drain.

The localised recharge from the shallow drain is largely responsible for the inconsistent recharge responses of the bores in this transect. In winter, bores 009, 010 and 011 respond more slowly and less than bores 007 and 008 on the valley floor. In winter the pressure head reverses, the area becomes a recharge site with the watertable higher than the groundwater level.

The drain appears to have negligible effect on the watertable beyond 20 m from the channel, as in transect 1. In-situ rainfall recharge, evaporation and low hydraulic conductivity appear to dominate the watertable movement. After drain construction, the area appeared to remain a groundwater discharge site except during those periods of shallow drain leakage, when it becomes a recharge site.

Profiles presented in Figure 35 show the watertable on four dates. For watertable profile 11 June 2004 the pre-drain levels are responding to the localised recharge from the shallow drain. By the 10 July 2004 the watertable had flattened out and the water level in 007 was falling, possibly in response to the drain. By 26 May 2005, the watertable had fallen to its lowest level, rising through the winter of 2005 and peaking in response to the 140 mm of early 2006 rainfall on 21 March 2006.

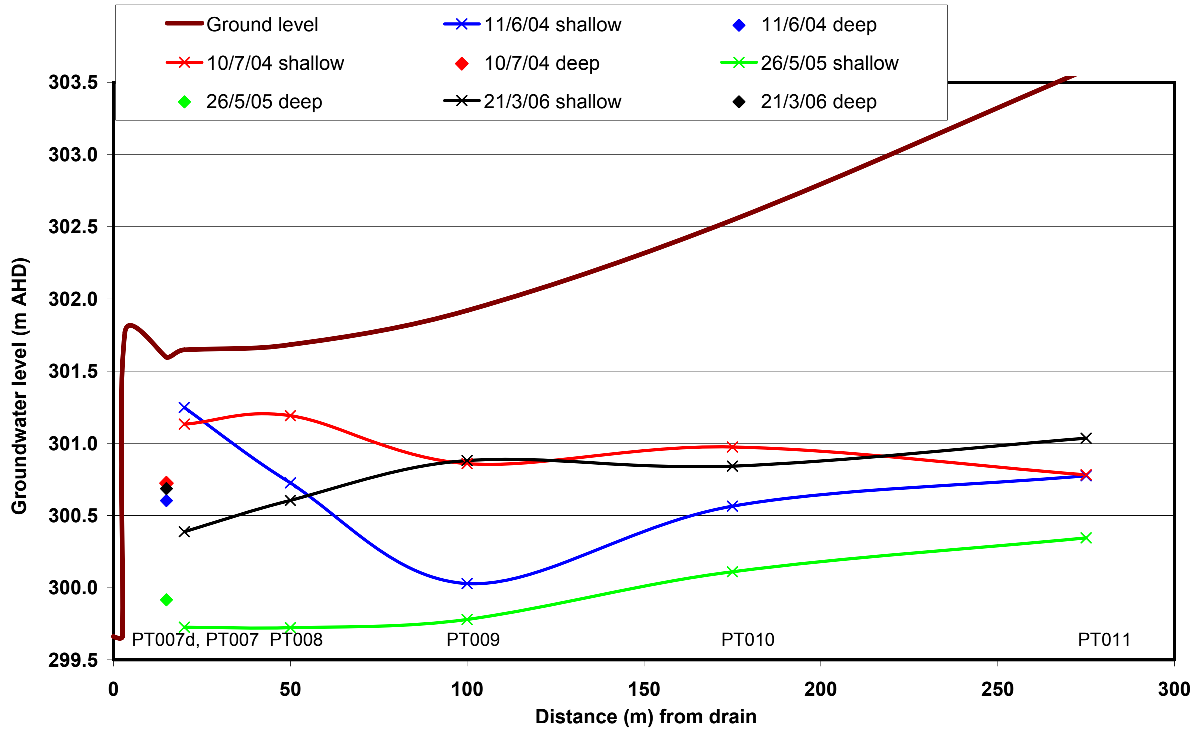

Figure 35 Transect 2 watertable profiles

Appendix A.4 Transect 3

These hydrographs show a relatively flat watertable along the transect reflecting its orientation partially across rather than perpendicular to the contour of the land. Transect 3 bores numbered 012–017 are towards the upstream end of the drainage scheme and perpendicular to a 600 m long tributary extending from the main drain towards a footslope seepage area to the north (Fig. 10).

Topsoils are pale grey loamy sands and sandy clay loam and the surrounding plant cover included barley grass, some rye grass and volunteer halophytes. A trial barley crop was established at this site during 2006 (Bell et al. 2009). The hydraulic conductivity in 013 was 0.056 m/d. The estimated hydraulic conductivity from inflow into an open pit dug at the drain alignment was 0.02 m/d.. The paddock surrounding transect 3 was probably the last of the salt affected land alongside the drain to be abandoned from dryland cropping, about eight years prior to drainage.

Construction of the drain past transect 3 on 9 July 2004 coincided with the onset of heavy winter rainfall (Fig. 3). The subsequent recharge in combination with possible drainage effects confused the analysis of the initial watertable drawdown caused by the drain. Five days after construction the drain lowered the watertable to a distance of 20 m (012), and later, possibly to 50 m (013) (Fig. 36).

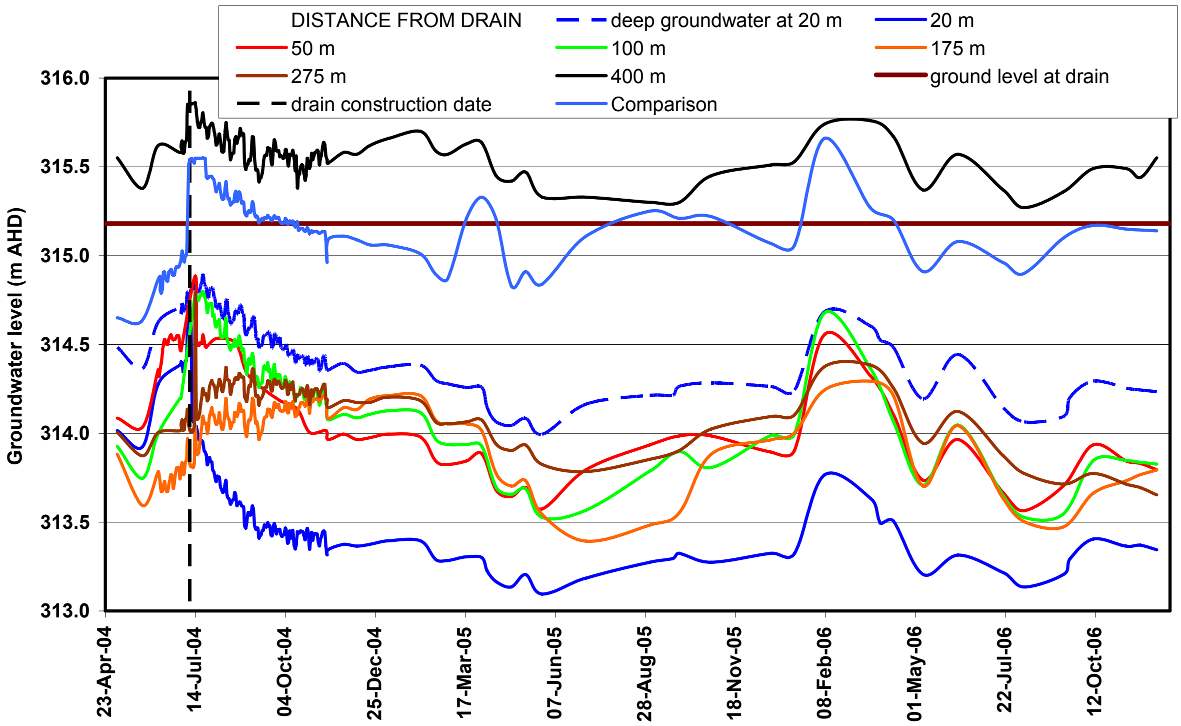

Figure 36 Transect 3 hydrographs (bores 012–017) and comparison bore 027

The water level at 20 m fell sharply by 0.6 m within the five days following construction. By November 2004 the water level was 0.6 m below its May 2004 level at the end of a dry summer (Fig. 36). On 26 May 2005 the water level fell to its minimum (313.10 m AHD), 0.8 m below 313.93 m on 27 May 2004. The water levels of the comparison bore 027 fell by about half these amounts.

Beyond 20 m, possible watertable responses to the drain were less obvious. From 14 July 2004–3 August 2004, the water level at 50 m (013) fell by 0.5 m, and, at 100 m (014), by about 0.2 m (Fig. 37). The lack of movement at 175 m (015) supports the possibility of a drain effect extending 100–175 m from the channel, if it were not for the similar watertable changes also at 275 and 400 m (0016 and 0017).

Bores responses at increasing distances are inconclusive: the watertable appears to respond sometimes at 100 m but not at 50 m, pointing to local recharge and evaporative effects. In the two and a half years of post-drain monitoring, levels in 012, 013 and 014 (50–100 m) fell well below 27 May 2004 pre-drain levels. The two lowest measurements were on 26 May 2005 and 8 August 2006 (Fig. 37). The drain did not appear to affect the watertable beyond 175 m. The water levels in 015, 016 and 017 (175–400 m) rose in the week after construction and remained within a much narrower range of fluctuation than those closer to the drain.

The groundwater in the deep bore 012d showed no initial response to the drain but started to decline 10 days later so it is not clear if this was a response to the drain (Fig. 36). The groundwater head was consistently above the watertable (012) with the separation between the two being about 0.5 m pre-drain increasing to 1 m post-drain.

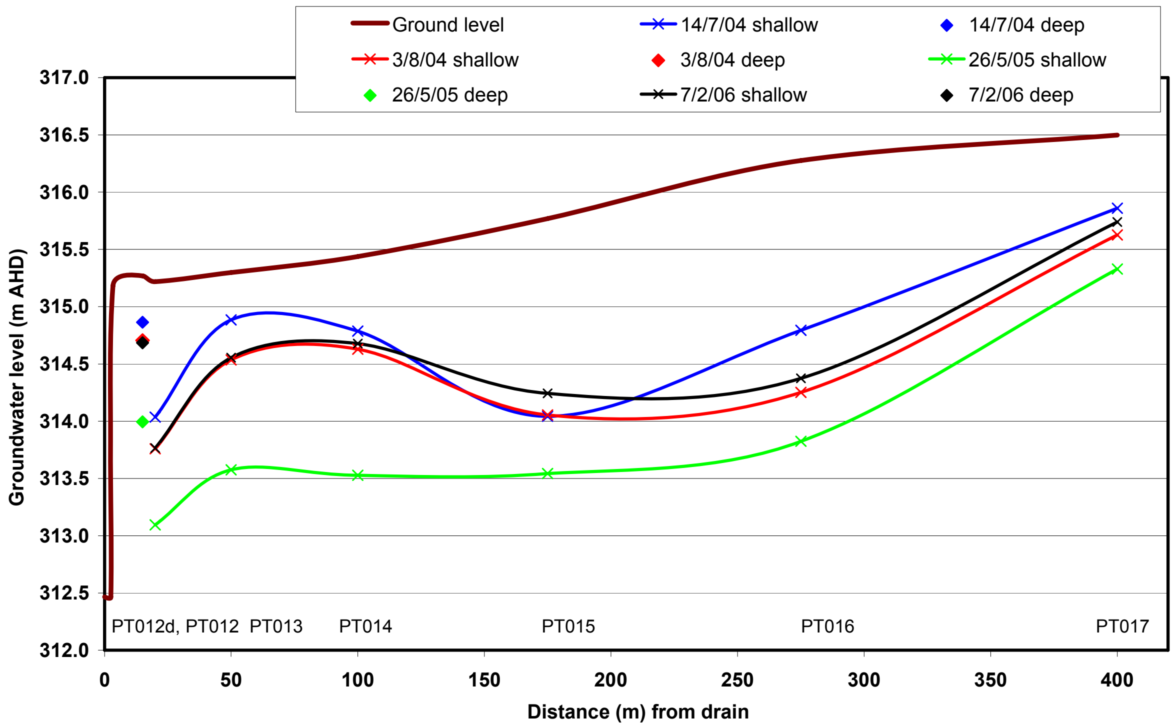

Figure 37 Transect 3 watertable profiles

Being orientated partially across the contour (Appendix CD 3.1) interpretation of groundwater flow being along transect 3 directly from right to left towards the drain in Figure 37 is inappropriate. If the watertable approximates that of the land surface, groundwater flow will in fact approach the drain at a 45° angle. The result is that theoretically groundwater moving past 015 (175 m) for example could be entering the drain 250 m further downstream.

Appendix A.5 Transect 4

Transect 4 bores 018–023 are 200 m from the end of the main drain (Fig. 10). This land had not been cropped for more than eight years before drain construction due to salinity. The sparse (~ 30%) plant cover was barley grass interspersed with volunteer halophytes (Fig. 12). Topsoils are grey sandy clay loam. The hydraulic conductivity measured in 019 was 0.015 m/d. Measurements on water seeping into an open at the drain alignment gave an estimated hydraulic conductivity of 0.017 m/d.

Drain construction reached transect 4 on 8 July 2004 coinciding with 21 mm of rainfall on the day and 15 mm on the previous day (Fig. 3). Although groundwater rose in response to the rainfall the rises were less than for the other transects and comparison bore 027. The watertable rose by about 0.25 m at 275 m from the drain, and by 0.5 m at 400 m (Fig. 38). Watertables in most of the other transect and comparison bores rose by 0.5 m or more in response to this onset of winter rainfall.

Groundwater pumping by the landholder after February 2005 and January 2006 noticeably affected the water levels at 175 and 275 m (021 & 022) and possibly other bores too (Fig. 38). The 28 m deep pumping bore was about 200 m south-east of the outer end of the transect. The hydrographs were variable and inconsistent after January 2005 and any ongoing drain effect on the watertable could not be determined. The pre-pumping May 2004–January 2005 hydrographs unaffected by pumping were used to assess impact.

At construction, water levels rose by 0.23 and 0.37 m respectively at 20 and 50 m from the drain. The water level closest to the drain was 0.14 m lower than at 50 m (Fig. 38). The newly dug drain may have enabled drainage of the recharging groundwater, preventing the level at 20 m (018) from reaching the peak response seen at 50 m (019). Between 8–18 July water levels in 018 and 019 fell respectively 0.17 and 0.27 m.

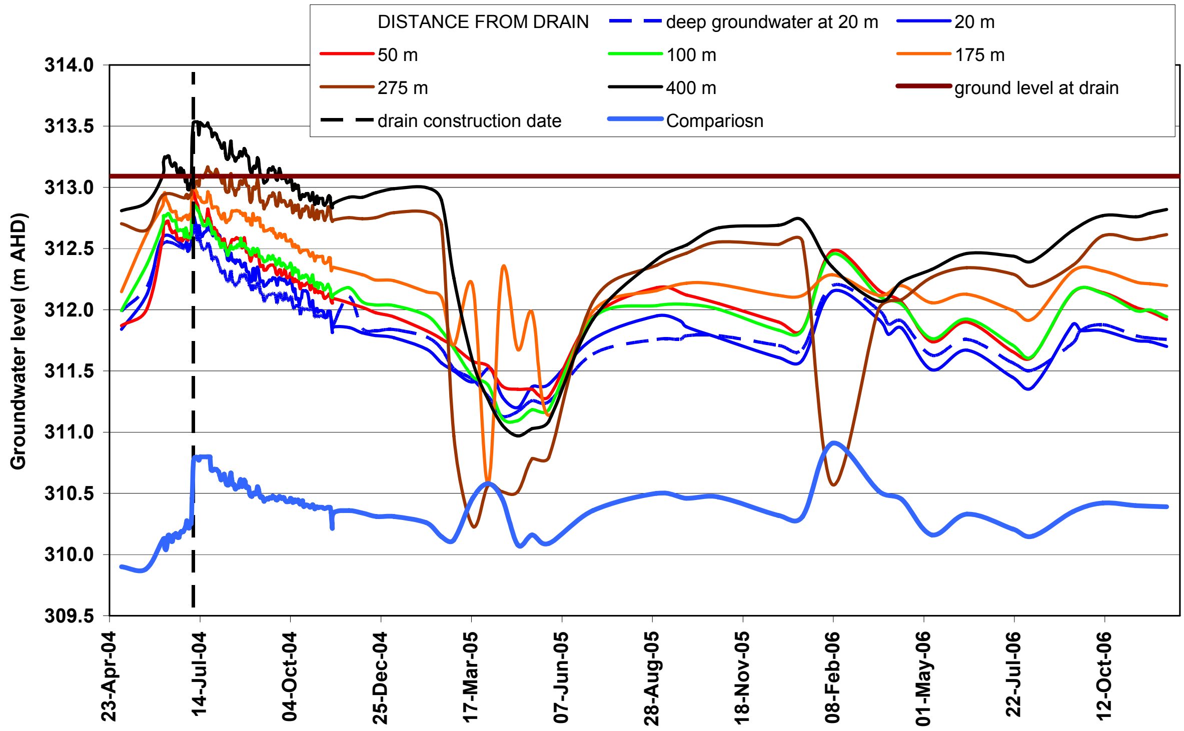

Figure 38 Transect 4 hydrographs (bores 018–023) including comparison bore 027

Watertable profiles show the pre-drainage (30 June 2004) peak watertable after drain construction (15 July 2004) and the return of the watertable close to the pre-drainage condition (3 August 2004) (Fig. 39). The lowest watertable (4 February 2005) was the last reliable measurement before levels were affected by groundwater pumping. Whilst high, the watertable approximates the land surface. As the watertable falls the rate of recession increased within 175 m of the drain (30 August 2004). By 4 February 2005 the watertable up to 175 m had fallen a further 0.5 m while beyond 275 m the level remained largely unchanged.

The drain has clearly affected the watertable as far as or just beyond 019 (50 m) where the watertable gradient increases near the channel (Fig. 39). The reduction in watertable height to about 175 m may also be a drain effect, but, considering the soil permeability and results from the other transects, may also be a groundwater evaporation effect.

The transect 4 deep bore (018d) was established to about 25 m depth. Even during the groundwater pumping, there was usually less than 0.1 m separation between the hydrographs of the groundwater and the overlying watertable. (Fig. 38). The groundwater showed no initial response to the drain but levels started to fall about two weeks later. Throughout the monitoring there continued to be a two-week lag between changes in the watertable and similar responses of the groundwater.

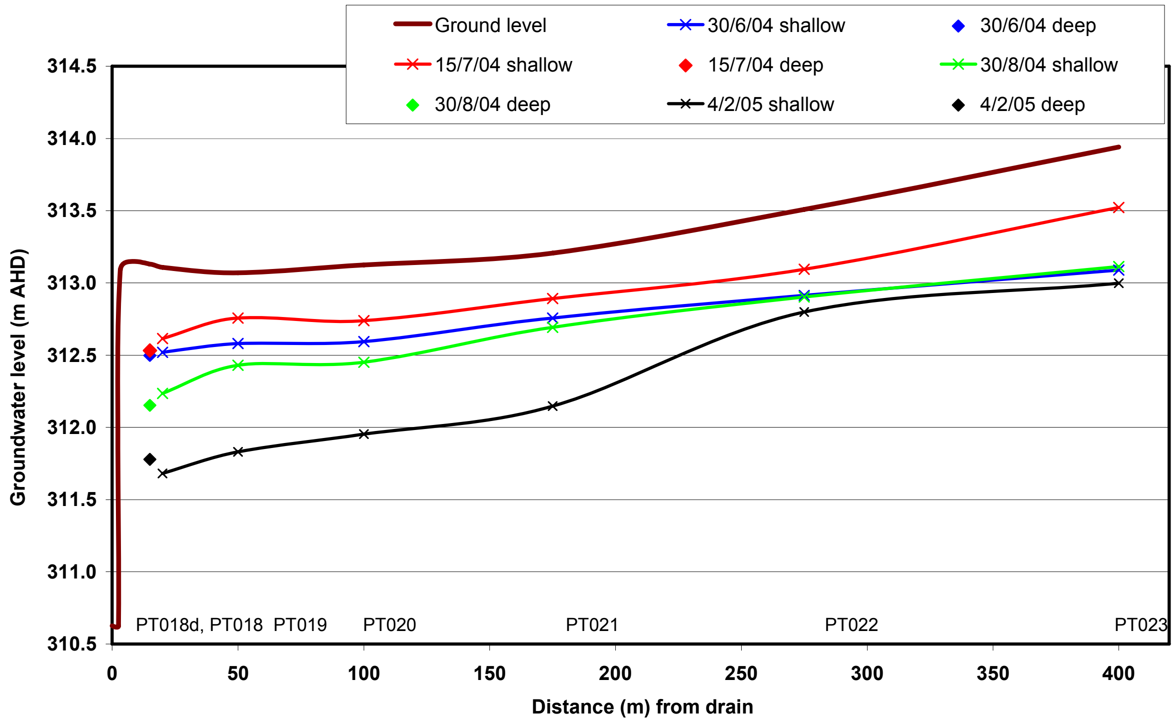

Figure 39 Transect 4 watertable profiles (bores 018–023)

Appendix B Water balance methodology

Appendix B.1 Bounded and unbounded drains

The Pithara drainage scheme consists of single channels constructed at or close to the lowest alignment along the valley floors. As such it has an unbounded groundwater catchment with no identifiable groundwater boundary within reasonable proximity. The concept of defined groundwater boundaries is critical in understanding the natural behaviours of groundwaters and induced responses to drainage. Hydrological boundaries are not necessarily impervious layers or walls confining the groundwater, rather they may be geometric surfaces and/or groundwater equipotential lines (Ritzema 1994).

Watertable management traditionally uses agricultural drainage with two or more parallel drains discharging into a common collector drain. The spacing between them is calculated to maintain the watertable below the required height, based on aquifer parameters and expected recharge. Because groundwater migrates towards the closest drain the watertable forms planes of symmetry around each drain and the highest point of the watertable marks the groundwater boundary between the drains (Fig. 40). These boundaries define the groundwater catchment for each drain.

The extent of each plane and the depths and so the volumes of recharge within the groundwater catchments are readily estimated. In designing a scheme the aim is to balance the peak volume of water to be drained from the groundwater catchment against the efficiency of the drain. If the expected recharge volume exceeds the rate of drainage the drain ‘spacing’ is reduced. Reducing the drain spacing has the effect of reducing the individual groundwater catchment for each drain, thereby reducing the volume of recharge to be drained.

In effect, the catchment area between the drains is ‘engineered’ by adjusting the spacing to enable drainage of the recharge within a specified time. The volume or depth removed over time is the ‘drainage rate’. The distance across each drain or the spacing between them from where the groundwater flows into the drains is usually referred to as the zone of influence (ZOI). If the groundwater flow from the ZOI is sufficient to lower or control the watertable between the drains, the ZOI could also be referred to as the corresponding watertable zone of influence (WT-ZOI). If the drain design so successfully controls the watertable that the WT-ZOI could be cropped, the land area would also fall within the zone of drainage benefit (ZOB) of the drain (Fig. 40).

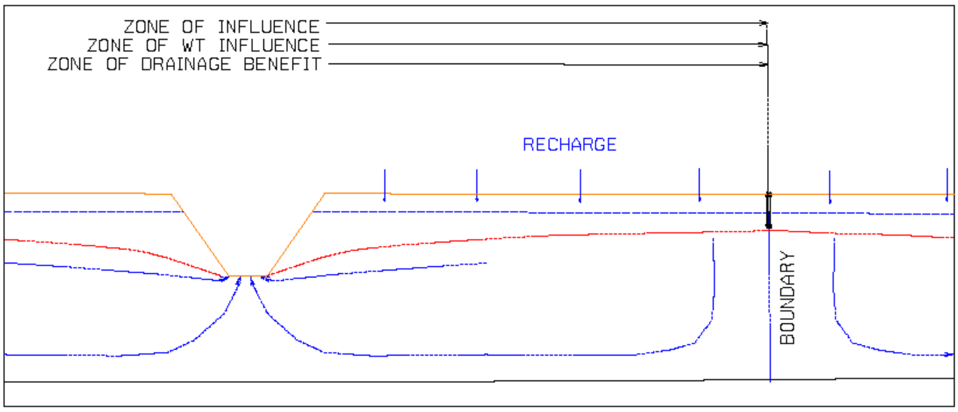

Figure 40 Watertable control by bounded drains

A drain is unbounded when the volume of recharge to its groundwater catchment continuously exceeds its drainage rate such that it is unable to lower or control the watertable. Essentially, the groundwater catchment is too big for the unbounded drain so yields groundwater volumes far in excess of drainage efficiency thereby reducing its capacity to lower the watertable.

In comparison to the bounded drain the ZOI of an unbounded drain more appropriately reflects the distance from which groundwater can migrate towards the drain without necessarily resulting in noticeable or significant watertable control (Fig. 41). This distance may be hundreds of metres, especially if the land alongside the drain is elevated and from which groundwater may naturally originate.

Closer to the drain the drainage rate could exceed the rate of local recharge and groundwater discharge from the ZOI, and the watertable level may fall. The identifiable extent of watertable reduction or control could be equated with the WT-ZOI of the bounded drain. If the watertable is lowered sufficiently within the WT-ZOI, land may be recovered for its intended purpose – in the case of Pithara, cereal cropping. The extent of this adequate watertable control is the ZOB.

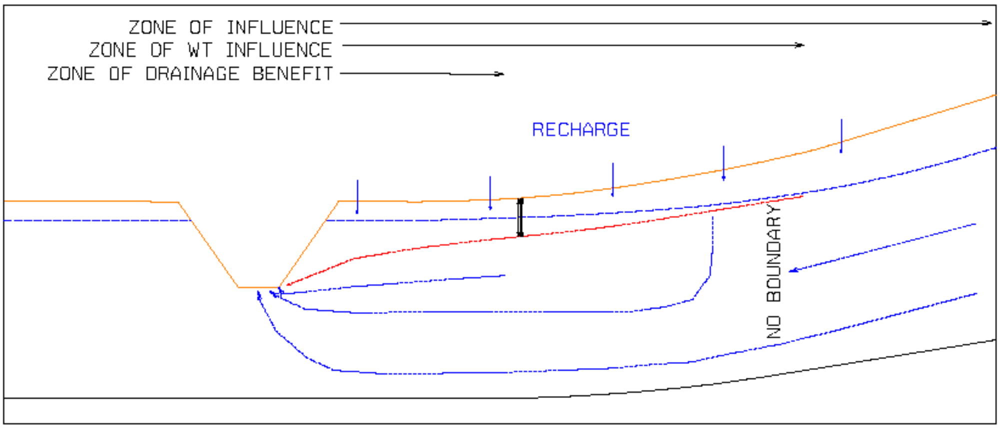

Figure 41 Watertable response to an unbounded drain

Where the groundwater catchment is designed to be proportionally correct in relation to the drainage rate of the bounded drain, recharge is drained and watertable rises are controlled. When the watertable rises the head between the watertable and drain floor increases driving a corresponding increase in drain discharge. As the watertable falls the decreasing head causes decreasing drain discharge until both the head difference and drain discharge can become zero.

If the groundwater catchment is unbounded, recharge can exceed the drainage rate at any time and the resultant watertable rise can 'swamp' any watertable control previously achieved. So watertable control provided by unbounded drains is more susceptible to climatic variability than control provided by bounded drains. In dry seasons the recovery of saline land by the unbounded drain could be quickly undone by wetter seasons or even major rainfall events.

Unbounded drains tend to produce sustained and uniform discharges due to the large supply of available groundwater. The discharge pattern reflects the relatively uniform head conditions that persist between the channel floor and the uncontrolled watertables along side.

Appendix B.2 Drain inflows, discharges and loads

Salt mass balances were used to separate groundwater inflows from the other sources of inflows and losses. Groundwater inflow is water that has seeped through the walls and floor of the channel from the surrounding aquifer. Once separated from the other variables that affect drain discharge the relationship between groundwater levels alongside the drain and inflow rates can be further explored.

The salt mass balance approach is based on the assumption that groundwater inflow is the source of salt inflow to the drain and that its salinity remains fairly uniform (Section 4.2). The uniformity of the salinity is underpinned by the apparent relationship between drain discharge salinity and that of the deeper groundwater (Section 4.5).

Temporary salt entrainment and displacement from the channel caused some unacceptably high and low calculated daily groundwater inflow rates (Appendix CD 5.2a & 5.2b). Salt which accumulated in the drain channel by evapoconcentration was mobilised by subsequent rainfall and runoff. The brief high salt loads at the drain outlet were translated by salt mass balance as high groundwater inflows. The freshening of the drain flow immediately following rainfall had the reverse effect. Because these inaccuracies are inherent from the daily results, inflow and other water balance results were aggregated and reported monthly.

In addition to groundwater inflow, the water balance of the upstream drain channel included accessions from rainfall and runoff and losses to evaporation (Fig. 42). These climatic variables affected the composition of the drain flow while in the channel. Without evaporation, rainfall and runoff, drain discharge would be of equivalent volume and quality to groundwater inflow. In all but two months, groundwater inflows were greater than total discharges because evaporative losses exceeded contributions from rainfall and runoff. Rainfall and runoff were equivalent in volume to about 17.5% of evaporative loss during the 30-month water balance.

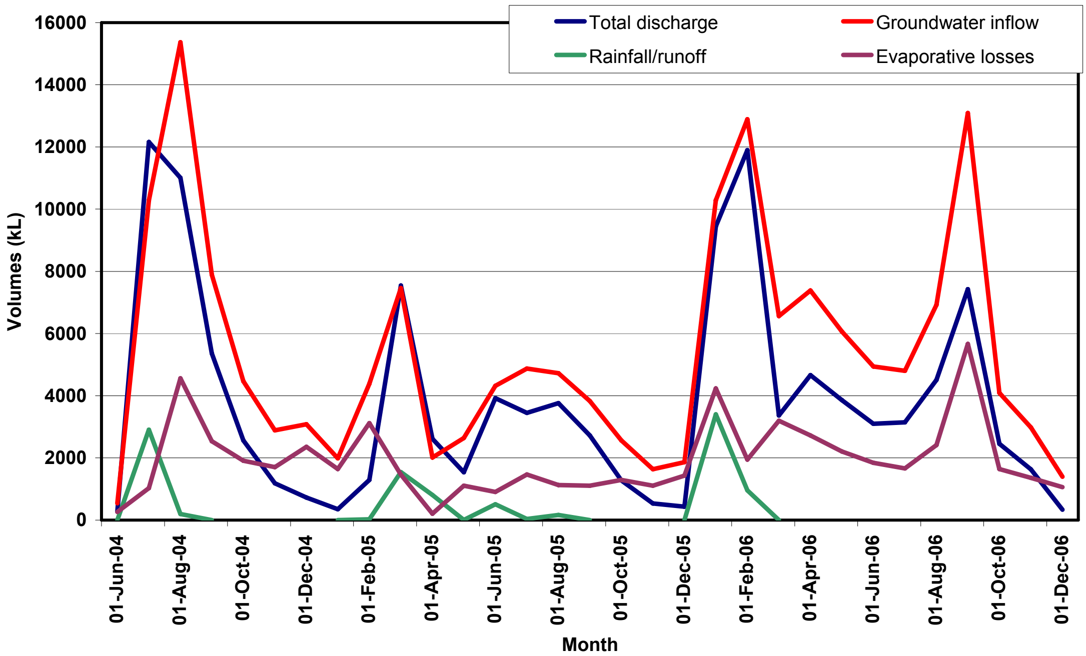

Figure 42 Monthly water balance of the upstream channel

The calibration salinity of groundwater inflow to the upstream drain was about 21 300 mg/L, calculated from the mean of the shallow and deep groundwater salinities from the transect bores at 20 m from the drain. These values were used to best represent the likely salinity of the underlying groundwaters seeping into the drain. The results of the salt mass balance gave 152 000 kL of groundwater inflow to the upstream drain (Appendix CD 5.2a).

The in-channel water balance of the downstream drain was more influenced by rainfall and runoff than the upstream drain (Fig. 43). Total discharge was dominated by rainfall and runoff in February 2006. Groundwater inflow almost stopped during the summers (December) of 2005 and 2006.

The small summer groundwater inflows do not accurately represent what is happening in the drain at these times. The calculated groundwater inflow from salt mass balance is based on the measured discharge and salinity at the gauging station. Because groundwater evaporated in the excavated creek between the deep upstream drains and the gauging station the summer groundwater flows were often not measured because flow did not reach downstream. This was known to happen because about 2840 kL of discharge from the upstream gauging station also did not reach the downstream gauging station during summer.

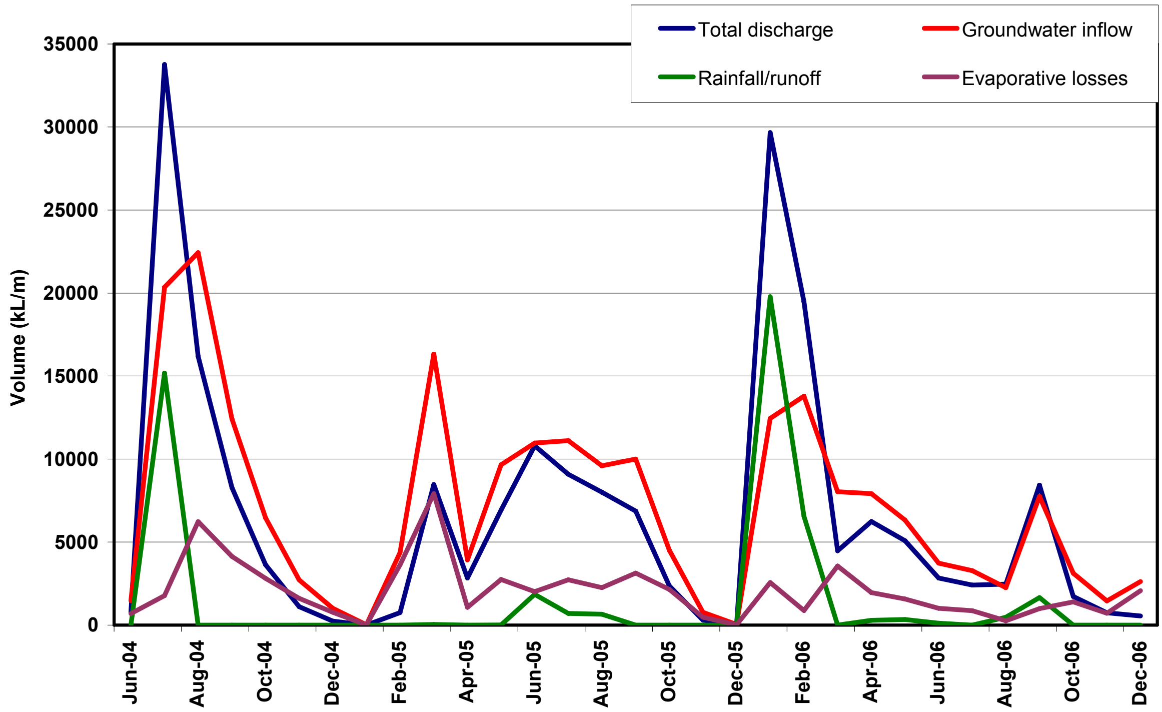

Figure 43 Monthly water balance of the downstream channel (less upstream inflow)

A groundwater inflow calibration salinity of 28 000 mg/L was estimated for the downstream drain from surrounding bores. The average daily flow-weighted salinity of the groundwater component of total discharge at the drain outlet was 86 200 mg/L, a three-fold increase. Groundwater inflow to the downstream drains was calculated by salt mass balance to be 221 000 kL (Appendix CD 5.2b).

The close relationship between monthly groundwater inflows to the upstream drain and AMRR (Fig. 44) produced a correlation of 0.7. For the downstream drain the poorer correlation value of 0.35 is a reflection of the physical factors that complicated the estimation of groundwater inflows for this drain. The most obvious influencing factors mentioned above include the drain being open to surface runoff, the effects of the excavated creek section and inflows from the upstream drain. These caused large variations in the delivery of discharge and salt to the gauging station that were transferred through to some inaccuracies in the salt balance approach.

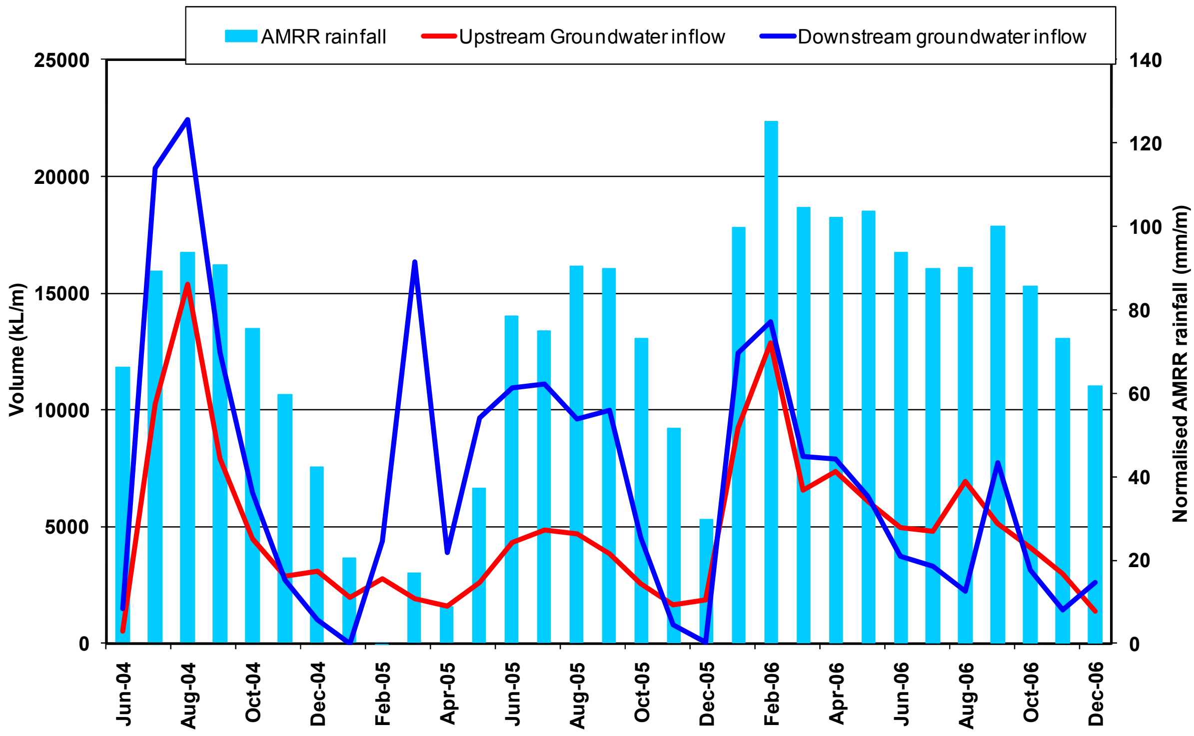

Figure 44 Upstream and downstream groundwater inflows and normalised AMRR

Appendix B.3 Groundwater heads and drain discharge

To demonstrate that a drain is removing groundwater and affecting the watertable there needs to be a sound relationship between groundwater heads and groundwater inflow to the drain (Appendix B.1). There were only enough bores in the transects to explore this relationship for the upstream drain. Because the relationship is one between groundwater head and groundwater inflows, groundwater levels are expressed as height or head above the drain channel floor.

From construction, groundwater inflow increased in response to the increasing length of channel and the drainage of accumulated groundwater from the upper part of the aquifer (Appendix B.4). At 62 days from the start of digging the inflow rate peaked at 545 kL/d (Fig. 45) (Appendix B.2). Construction of the drain took 31 days and watertables were all falling by day 62.

From day 62 the relationship between higher watertable heads and higher inflows, and vice versa is visible. This was as expected because the relationships between watertable levels and AMRR, and groundwater inflow and AMRR were previously established (Appendix A & Fig. 44), hence a relationship must exist between heads and inflows. Particularly in the Decembers of 2004, 2005 and 2006 it was quite evident that average watertable heads along the transects were at similar levels, resulting in similar groundwater inflows, around 60 kL/d, to the drain.

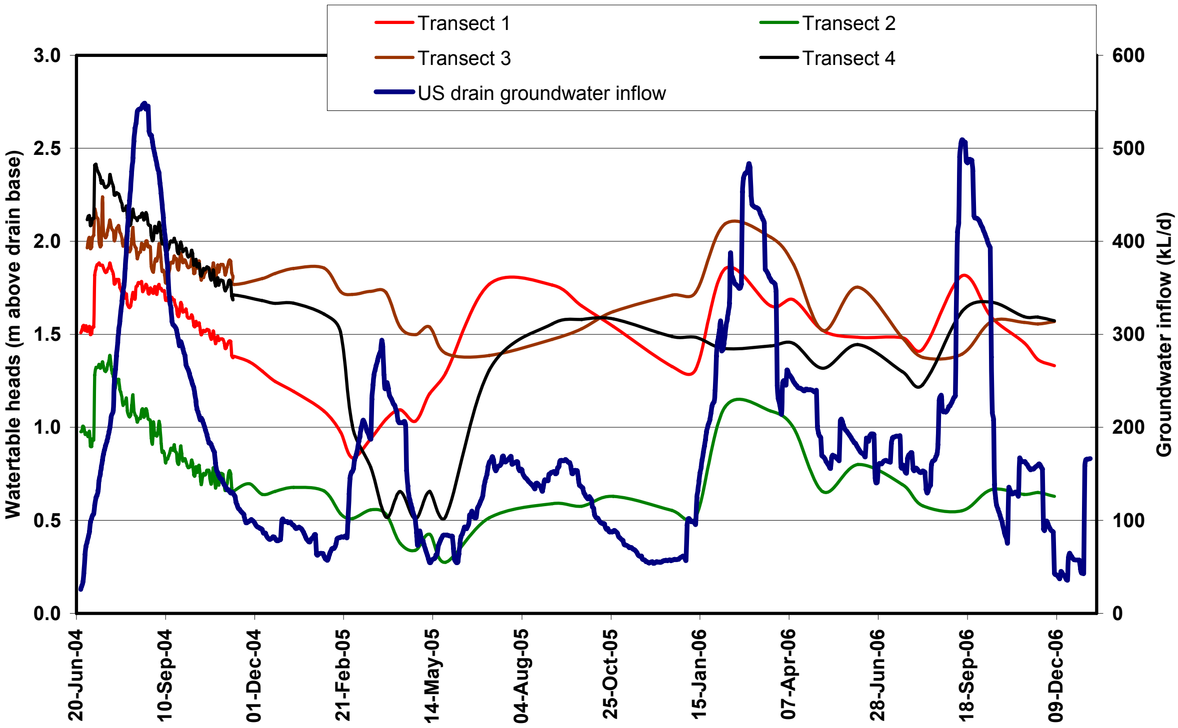

Figure 45 Watertable heads above the drain floor for transects 1–4 compared to groundwater inflows

There is a difference between having both the watertable levels and the drain responding to the same climatic variables, and the drain 'controlling' the watertable level. If the drain controlled the watertable level more lowering effect on the watertable nearer the drain than further away would be expected. As a result there should be a closer relationship between the watertable heads nearer the drain and groundwater inflows because the fluctuating watertable levels nearer the drain contribute more directly to groundwater inflows than those further away (Cox & Tetlow in press).

Correlation coefficients were used to explore the relationships between watertable heads along the transects and groundwater inflows to the drain. Expected results would produce high correlations close to the drain diminishing to nearly zero at distance. The zero or low correlation represents the distance at which the drain has no effect on the watertable, equating with the outer limit of the WT-ZOI (Appendix B.1).

Transect 1 produced a result closest to expectations (Fig. 46) but tends to show that the WT-ZOI theoretically extends approximately 1000 m on each side of the drain (from trend analysis). The trend of the transect 4 correlation is also orientated correctly in relation to the drain, indicating a closer relationship between the watertable head and groundwater inflow nearer the drain. There appears to be little or no correlation between the watertable heads and groundwater inflow for transects 2 and 3 with the linear trends in these correlations reducing towards the drain to near zero values.

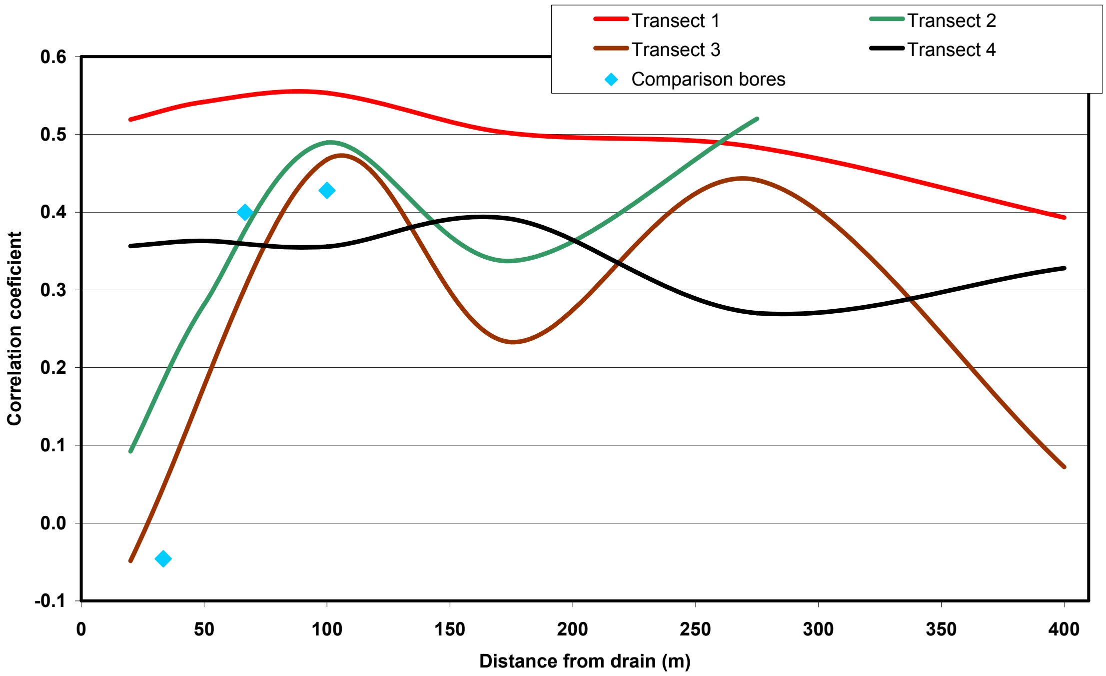

Figure 46 Correlations between watertable heads along the transects and groundwater inflows

The relationship between average monthly watertable heads and total monthly groundwater inflows produced a correlation of 0.6 for the entire 31 months of monitoring. For the groundwater heads the correlation with groundwater inflows was above 0.7. There is a correlation of above 0.9 for the relationship between groundwater heads and watertable levels.

The moderate and best correlation between groundwater heads and inflows appears to indicate that drain flows are primarily dominated by deeper groundwater rise. This observation is supported by the nearly continuous positive head differences between the groundwaters and watertables measured alongside the drain (Appendix A). A high correlation value of 0.9 for the average monthly groundwater heads and watertable levels also reveals connectivity between the two. Combined with the positive head difference the correlations shows the watertable level is significantly influenced by that of the underlying groundwater heads.

Appendix B.4 Aquifer storage change

Fluctuations in the watertable alongside the drain reflect changes in aquifer storage. Rising watertables show increasing or positive changes in storage while falling or draining watertables show decreasing or negative changes.

Post-drain changes in aquifer storage were estimated from the changes in watertable levels in the transect bores. Only the three bores within 100 m of the drain from each transect were used for the calculations. The results were then extrapolated to represent the changes in storage beneath whole 182 ha drainage site (Section 5.1).

Changes in the watertable cross sectional area in each transect were multiplied by a specific yield value of 0.01 (Smedema & Rycroft 1983) and 1 m width. Specific yield and the 1 m width allowed changes in watertable height to be expressed as changes in groundwater volume beneath each 1 m wide by 100 m length strip of land along each transect. The volume changes were totalled and reported monthly.

Positive and negative storage changes beneath the transects mostly coincided with each other, particularly where large changes were involved such as in February 2006 (Fig. 47). Extrapolated storage changes for the comparison bores (Appendix CD 5.4) mostly coincide with those occurring beneath the transects.

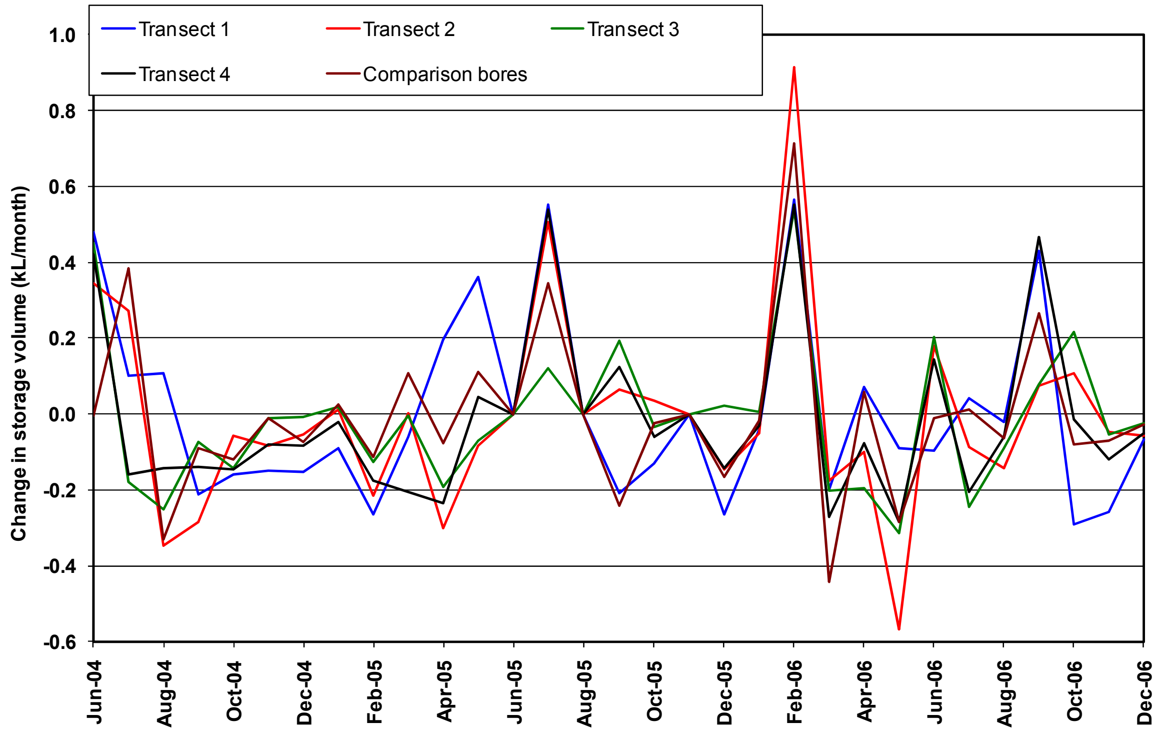

Figure 47 Monthly changes in groundwater volume beneath transects 1–4 and comparison bores

The average change in aquifer storage from the transects was extrapolated to the whole of the drainage site. When drain construction started in June 2004 the aquifer storage volume increased by about 7500 kL/month while groundwater inflow to the partially constructed drain was 500 kL/month (Fig. 48). After drain completion in July 2004 inflow had increased to 10 000 kL/month and aquifer storage change was zero. The aquifer storage continued to decline (-ve values) until the onset of rainfall in May 2005. Total aquifer storage reduced by 19 000 kL from July 2004–May 2005 by which time cumulative groundwater inflow was 63 000 kL.

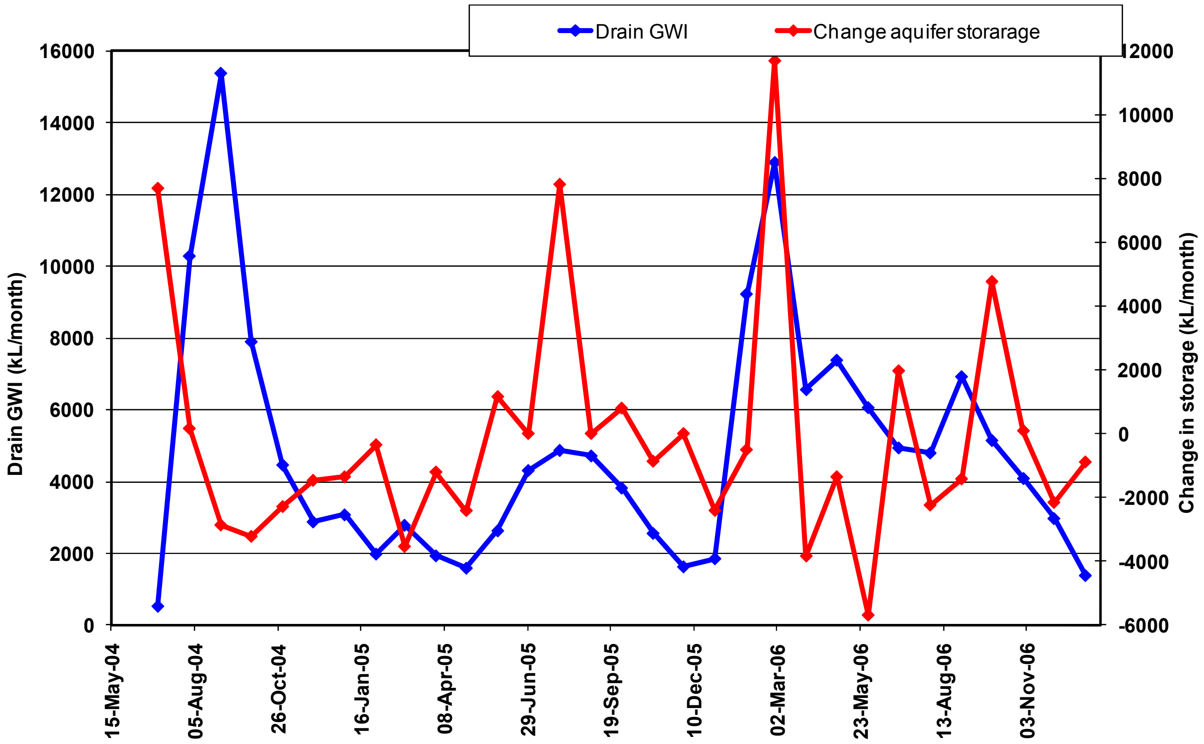

Figure 48 Aquifer storage volume changes beneath the drainage site and groundwater inflows

The addition of about 36 000 kL of groundwater was responsible for the increased aquifer storage during the post-drain monitoring period (Fig. 49). This was counteracted by the loss or drainage of 40 000 kL, leaving a net reduction in storage of about 4000 kL. Throughout the monitoring period the aquifer storage reduction was consistently about 25% of groundwater inflows to the drain.

The cumulative filling and draining hydrographs for aquifer storage reflect the mainly winter filling and summer draining pattern except during February 2006. The extreme rainfall that month caused a 12 000 kL increase in storage that was not drained until May–June that year (Fig. 49).

Although aquifer storage change is dynamic not all groundwaters migrating into the drain necessarily show in storage changes. Just as a river can flow without changes in water level, groundwaters can flow into the drain without causing changes in watertable levels as long as positive head conditions (Appendix B.3) and groundwater supply are maintained. Changes in aquifer storage are responses to changes in the rates of groundwater supply compared to the rates of drainage or other losses.

The frequency of the monitoring was only enough to calculate the storage change between the end of one month and the next. Watertables and aquifer storage could have increased or decreased to a greater extent than calculated with no difference in the final 4000 kL net reduction.

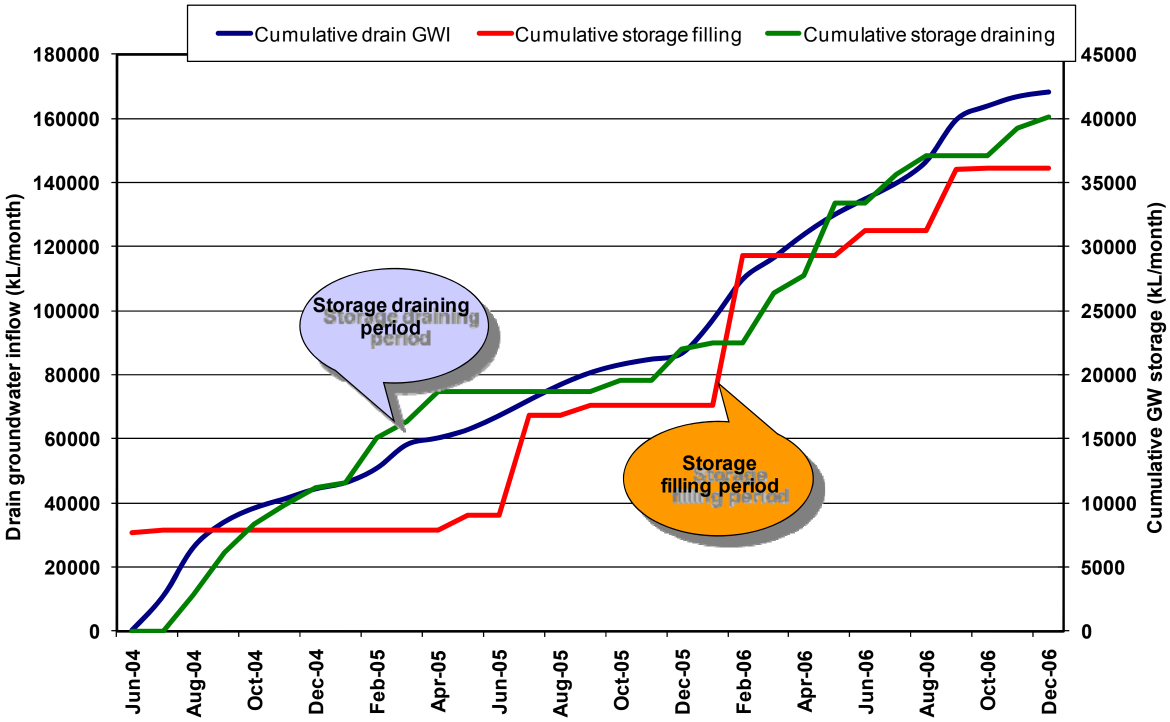

Figure 49 Cumulative monthly aquifer storage filling and drainage balances and groundwater inflows

Appendix B.5 Rainfall recharge

The average post-drain rain-fed recharge to the drainage site was calculated at 10.5% (78.5 mm) of the 650 mm of rainfall. Recharge was estimated using the specific yield technique (McFarlane et al.1987). Average watertable rises beneath each of the transects provided the watertable level data for the analysis. The daily water level measurements used were only available for the period of automated monitoring 11 June 2004 to 12 November 2004. A regression analysis of the rainfall–recharge relationship was used to extend the results for the duration of monitoring (Appendix CD 5.5).

The exponential relationship between rainfall and watertable rise produced R^2 values of about 0.7 (Transect 4) and as low as 0.05 (Transect 3). The extrapolated post-drain results ranged from 42 mm (6.5% of RF) for transect 1 to 95.4 mm (14.7% of RF) for transect 4, based on the July 2004–December 2006 rainfall (Fig. 3). The highest monthly recharge was 37.9 mm for transect 4, in response to 92 mm of rainfall for the month (Fig. 50). Averaging the transect recharge results and extrapolating the average value across the drainage site produced 125 000 kL of recharge.

This simplistic approach to recharge estimation belies the true complexity of the recharge processes and should at best be regarded as estimated recharge. Due to the short duration, limited range of available data and regression analysis coinciding with the drying period of the year, recharge at some transects could be underestimated. Ali (2004) used recharge rates as high as 13% of rainfall for predicting drain performance at Narembeen, in the eastern Wheatbelt. However, it was not indicated if this value was only representative of rain-fed recharge or was inclusive of aquifer discharge (Appendix B.6). Rain-fed recharge of 8.5–10% of rainfall was calculated from watertable and drain responses at Dumbleyung (Cox & Tetlow in press).

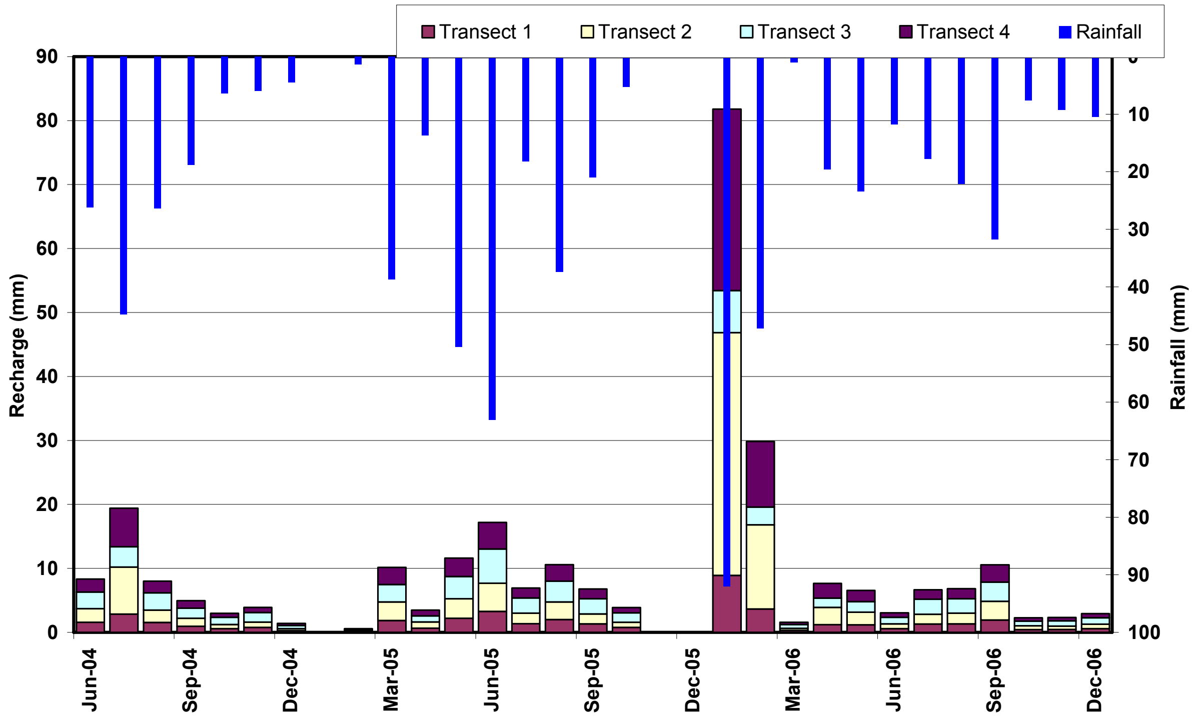

Figure 50 Rainfall and calculated recharge for transects 1–4

Appendix B.6 Aquifer discharge

The aquifer beneath the valley floor is essentially viewed as a continuum of the aquifer beneath the hillsides. This groundwater connection reinforces the assumption that a proportion of the groundwater within the valley floor originates from hillside aquifer discharge. In the context of the unbounded drain (Appendix B.1) aquifer discharge is viewed as groundwater originating from within the ZOI but outside of the WT-ZOI. This makes aquifer discharge a potentially significant component of the water balance of the Pithara drain site.

Aquifer discharge can appear to move large distances towards a drain because of the pre-existing hydraulic gradients that extend away from the drain and valley floor to beneath the footslopes and valley flanks. Once groundwater enters the valley floor aquifer flow paths towards the drain can become radial as well as horizontal. Radial groundwater flow can occur beneath the Pithara site because the ratio of aquifer thickness to drain depth is approximately 10:1–15:1, the drain is unbounded and the aquifer semi-confined.

The flow-net model (Fig. 51) demonstrates that 'groundwater streamlines' can develop extensive radial flow paths towards the drain in homogeneous aquifer conditions. The flow lines can penetrate the full thickness of the aquifer and can greatly enhance the lateral impact of the drain. Radial flow highlights the potential groundwater interconnectedness between the drain and valley flanks making it impossible to differentiate between local rain-fed recharge or aquifer discharge as the sources of groundwater inflow to the drains.

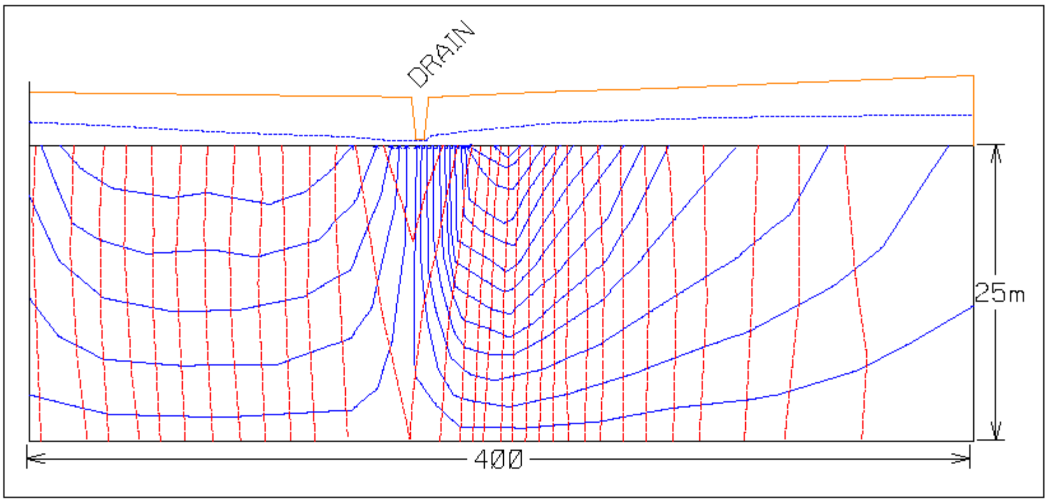

Figure 51 A flow-net in a homogeneous aquifer beneath a groundwater drain

The recharge and aquifer discharge components of the water balance of an unbounded drain cannot be accurately solved without using complex groundwater modelling. A large component of modelling would include calculating hillside aquifer discharge to the drainage site (Section 5.1) from its recharge and aquifer characteristics across the catchment.

An alternative approach to complex modelling is to calculate aquifer discharge into the drainage site using the Dupuit assumption (Ritzema 1994). This method can provide a reliable estimate of the rate of aquifer discharge into the drainage site through an arbitrarily placed 'porous' groundwater boundary (Section 5.1). For the upstream drain, the arbitrary boundary was placed 100 m from the drains (Fig. 26).

Using the Dupuit assumption, the aquifer discharge contribution to the drainage site is calculated from its discharge or outlet end through a porous boundary rather than at its recharge or inlet ends elsewhere in the catchment. The method applies the Darcy assumption (Ritzema 1994) to the slope of the watertable at a chosen point(s). The Dupuit assumption was applied to the intersections between the porous boundary around the perimeter of the drainage site and transect bore alignments.

In this method, aquifer discharge represents the relationship between groundwater heads outside the drainage site compared to the groundwater heads inside the drainage site. This is believed to be an accurate reflection of the head-driven process that partly controls aquifer discharge to the valley floor aquifer.

Aquifer discharge was calculated to contribute a total of about 100 000 kL into the drainage site from July 2004 to December 2006. For most of the time there was an inverse relationship between average monthly watertable heads within the drainage site and aquifer discharge (Fig. 52).

Aquifer discharge calculated from transect 2 produced the largest range of average monthly discharges from 0–1.1 kL/d per metre boundary length. Transect 2 extends from the drain up the sandplain valley flank so more strongly reflects the seasonal pulses of groundwater recharge migrating from the valley flanks to valley floor. Average monthly discharges for transects 1 and 4 were fairly uniform at around 0.2 and 0.04 kL/d/m, respectively. The discharge values for transect 3 were not used because the orientation of the transect across the contour caused irregular watertable profiles that resulted in erroneous results.

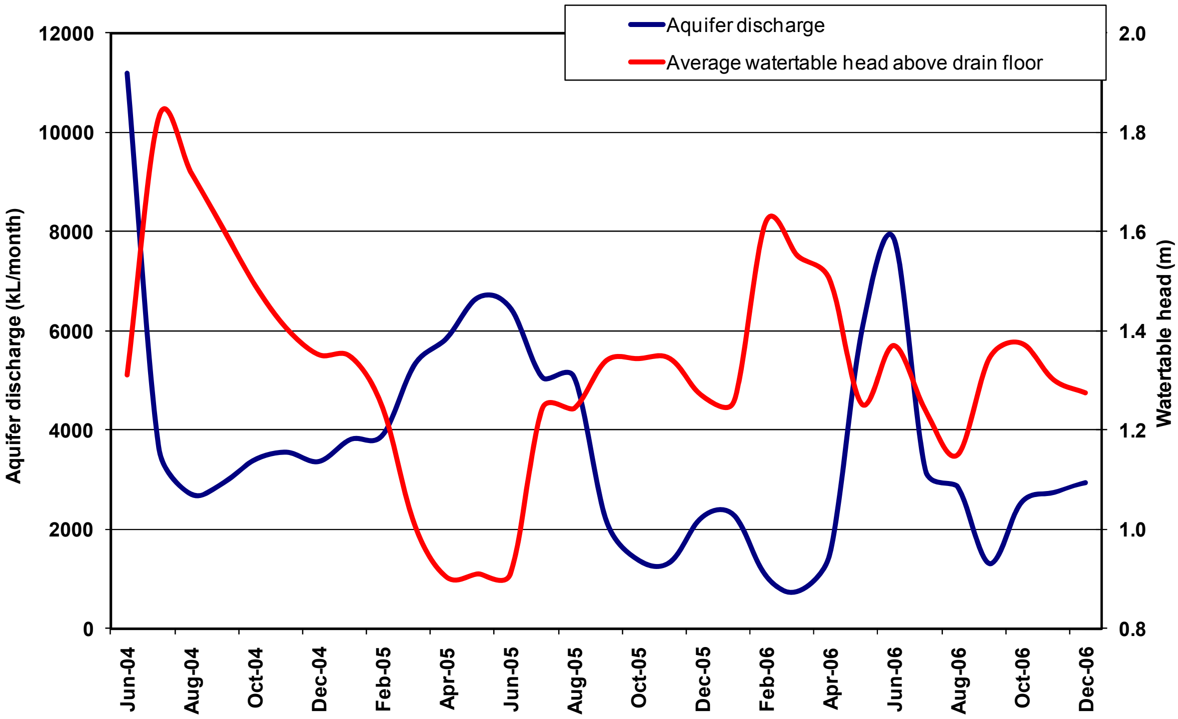

Figure 52 Inverse relationship between watertable height and aquifer discharge

The construction of the drain did not appear to increase the aquifer discharge rate into the drainage site. This is thought to be due to the relatively low transmissivity of the valley floor aquifer. Instead, potential aquifer discharge reduced noticeably at drain construction in response to the rain-fed watertable rise across the valley floor at the time (Appendix A). Reducing watertable heads within the drainage site compared to those outside allowed aquifer discharge to increase during the drier months to more than 6500 kL/month in May 2005 (Fig. 52).

Appendix B.7 Capillary discharge and groundwater evaporation

Capillary rise and subsequent evaporative loss showed the greatest potential to remove groundwater from the drainage site. Capillary rise was calculated to have had the potential to transport more than 350 000 kL of groundwater from the watertable to the soil surface during June 2004–December 2006. By comparison the 5401.8 mm potential evaporation (Section 2.1) had the capacity to evaporate nearly 1 000 000 kL from the site. On most days there was enough evaporative deficiency to evaporate all capillary rise as capillary discharge.

Capillary rise is the upward force lifting a water column due to the surface tension between the water and the circumference of the capillary (Smedema & Rycroft 1983). Capillary rise can lift groundwater to heights above the phreatic watertable, depending on the water tension in the root zone and the diameter and connectivity of the soil pores (capillaries). The upper extent of the capillary rise is the capillary fringe.

As the watertable rises the capillary fringe enters the root zone making the groundwater available for plant use. As the watertable keeps rising, the capillary fringe rises (theoretically) above the soil surface. From this height groundwater can be lost from the soil surface by capillary rise and evaporation. This is capillary discharge of the groundwater. The closer the watertable to the soil surface, the higher the potential capillary discharge.

The relationship between soil texture, watertable height and capillary rise (Fig. 53) has the 'Y' axis as the depth to the watertable and the 'X' axis as the rate of capillary rise at a surface with soil moisture pressure of 1600 kPa. For the silty loam soil type, capillary rise is near zero while the watertable is at 4.8 m below ground (Fig. 53). Capillary rise increases to 5 mm/d when the watertable is at 1.6 m below ground. In coarser-textured soils, the watertable can be closer to the surface before reaching the same levels of capillary rise as silty loam.

Capillary discharge has a very significant influence on controlling the pre-and post-drain watertable heights and groundwater discharge rates across the Pithara drainage site. The high rates of capillary discharge are substantiated by the accumulation of salts across the land surface of the valley floor and footslopes. As saline groundwater is transported to the surface and evaporated, the salts left behind accumulate on the surface and in the root zone (Fig. 54).

The rates of capillary rise were calculated from the fluctuating groundwater levels across the site using the coarse loamy sand representative soil type (Fig. 53). The transect bore results were further extrapolated to represent the potential volume of capillary rise across the upstream drainage site. Although any capillary rise might be assumed to become capillary discharge, this is not so. If the rate of capillary rise exceeds evaporative demand or if evaporative demand is satisfied by rainfall, the soil moisture pressure at the surface is reduced and so reduces the rate of capillary discharge. Because the capillary discharge rates have not been adjusted for these factors, capillary rise reflects only potential capillary discharge.

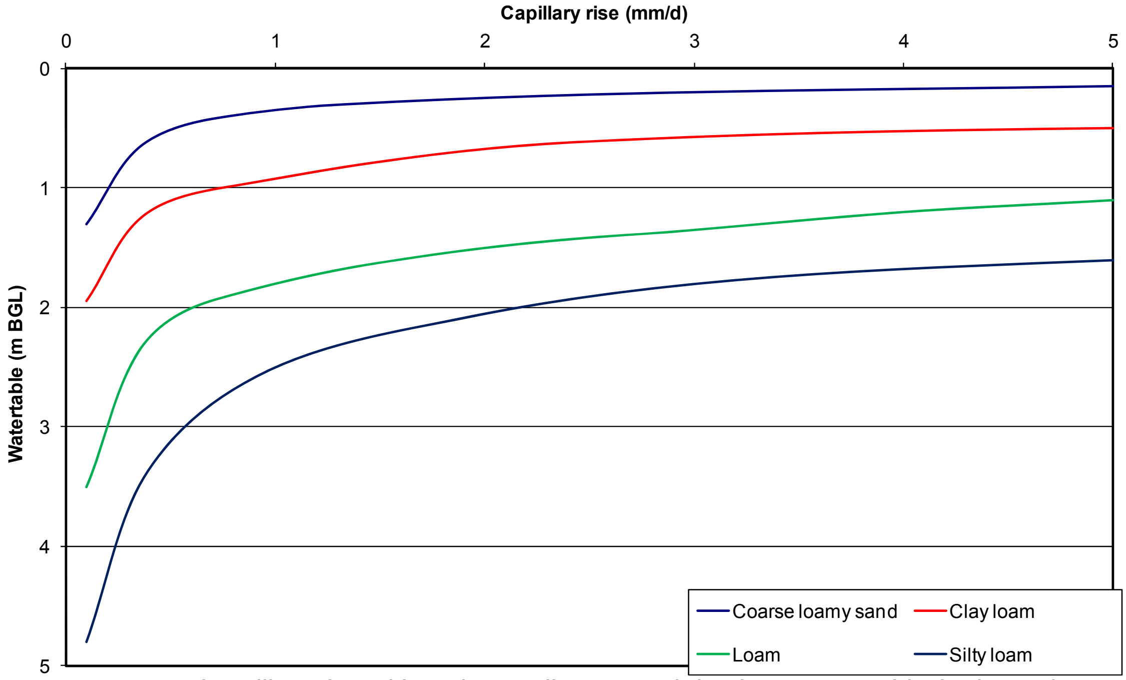

Figure 53 Rates of capillary rise with various soil types and depths to watertable (redrawn from Smedema & Rycroft 1983)

F*igure 54 Salt accumulated on the soil surface when depth to watertable is 0.8 m*

With watertables sometimes rising to within tens of centimetres of the soil surface, transect 1 showed the greatest range in potential capillary discharges equivalent to 5–50 mm/month (Appendix CD 5.7). Potential capillary discharge from transect 1 peaked at 50 mm in August 2005, coinciding with 37.4 mm of rainfall and 82.4 mm pan evaporation. With only an average evaporative deficiency of 45 mm for the month it was not possible for the true capillary discharge to have reached its potential 50 mm rate.

Fluctuations in potential capillary discharge appear most sensitive to watertable level fluctuations at around 1.4 m below ground level. This is a reflection of the steeper decline in capillary rise rate from the clay loam soils when the watertable falls below 1.4 m (Fig. 53). Reductions in average watertable levels below 1.4 m result in significantly lower capillary discharges with discharge approaching zero when the watertable nears 2.0 m (Fig. 55).

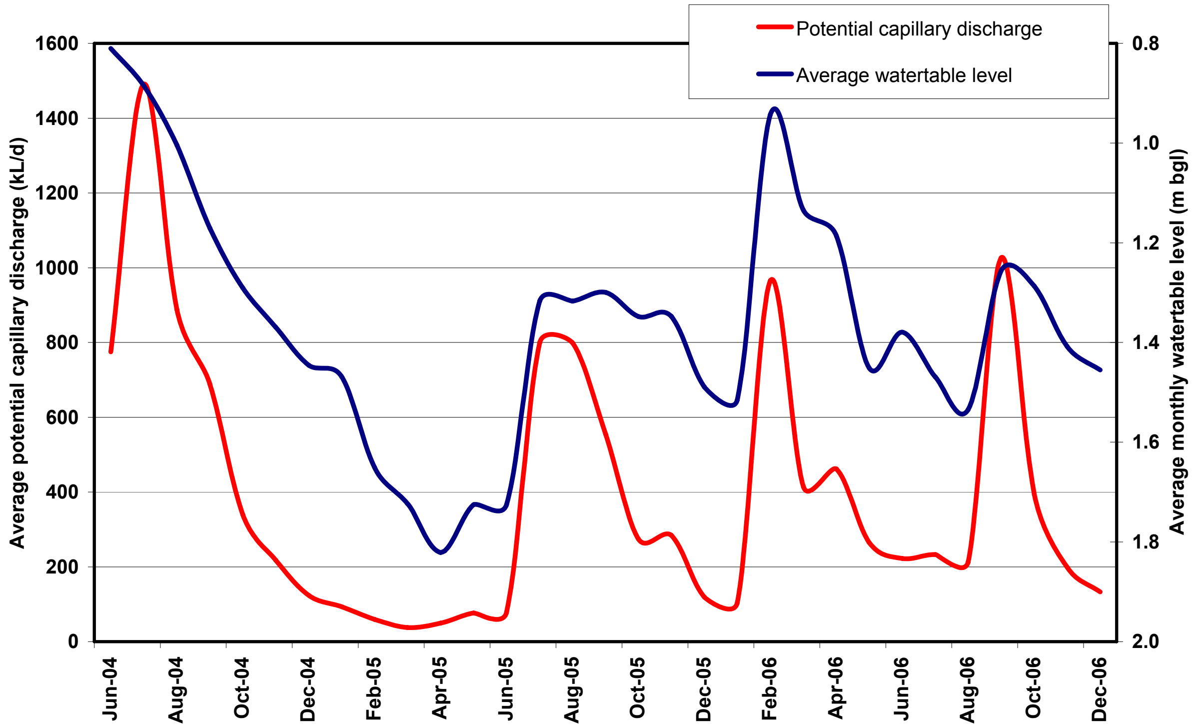

Figure 55 Monthly potential capillary discharges and average watertable depths

Glossary and abbreviations

AMRR	Accumulated monthly residual rainfall is the progressive accumulation of rainfall for each month less the average monthly rainfall for the period of analysis
Aquifer	A water-bearing soil layer that can store and transmit extractable volumes of water
Aquifer discharge	(A_D) The movement of groundwater into the drainage catchment (kL) from beneath surrounding elevated lands
Baseflow	Discharge from the drain that is derived from groundwater inflow (kL)
Batter	The inside edges of the drain channel that extend from the natural ground level down to the floor of the channel
Batter slope	The slope of the batter expressed as a ratio X:1, vertical to horizontal distance
Berm	The strip(s) of land between the top of the drain channel batter and inside toe of the levee bank
Bounded drain	A drain in a parallel scheme where each drain forms a groundwater boundary to another. When appropriately spaced, the zone of influence (ZOI), watertable zone of influence (WT-ZOI) and zones of benefit (WT-ZOB) are all equal and aquifer discharge is at or near zero
Capillary discharge	The loss of groundwater transported upwards by capillary rise, by evaporative processes from the soil surface (mm)
Capillary rise	The upward movement of water from the watertable caused by the adhesion of water to the soil and its resultant movement through the soil pores (mm/d)
Channel	The excavated part of the drain structure that conveys or intercepts water
Critical depth	(1) The depth below which a saline watertable must be maintained to meet the land use objectives. (2) The watertable depth at which capillary discharge is reduced to almost zero
Cross sectional area	(CSA) The area of a truncated end or section of a structure such as a drain channel (m^2)

De-silting	The mechanical removal of accumulated detached soil from a drain channel
Detention basin	A purpose-built reservoir for storing drain discharge
Discharge	The total volume of all water that flows from the outlet of a drain or drain section (kL)
Discharge hydrograph	A graph showing flow rate as a function of time at a given location
Drain structure	All the components of a drain: channel, berms and levees (if present)
Drainage catchment	(Drainage site) The land area surrounding the drain beneath which drawdown could or has occurred (ha)
Drainage coefficient	The discharge from a groundwater drainage system expressed as the depth of water removed within a certain time
Drawdown	A reduction in watertable height caused by the drainage of groundwater by a groundwater drain (see watertable zone of influence)
Drawdown profile	The drawdown measured along a transect perpendicular to a drain or other point of interest
Erosion	The removal of detached soil by rainfall, wind and moving water
Groundwater	Water within an aquifer below the watertable
Groundwater drain	An excavated channel that penetrates the aquifer for the purpose of draining groundwater
Groundwater discharge	The groundwater component of discharge from the drain outlet
Groundwater inflow	The movement of groundwater into the channel of a groundwater drain from the surrounding aquifer
Groundwater outflow	The movement of groundwater from the aquifer surrounding a groundwater drain into the channel. Groundwater outflow from the aquifer becomes groundwater inflow to the channel
Halophytes	Salt tolerant plants
Hectare	(ha) An area of 10 000 m^2

Hydraulic conductivity	(K or K_{sat}) A constant of proportionality in Darcy's Law defined as the volume of water that will move through the soil in unit time and unit hydraulic gradient through a unit area measured at right angles to the direction of flow (Ritzema 1994)
Hydraulic gradient	The slope of the watertable (m/m)
Kilolitre	1000 L or 1 m^3 (approx.) of water (kL)
Kilometre	1000 metres distance
Levee bank	A continuous mound of earth used to exclude or redirect runoff
Leveed drain	A groundwater drain with the channel completely enclosed within levee banks
Linear metre	(Lm) Measured distance along an alignment or the alignment of a structure
m AHD	Height in metres above the Australian Height datum taken as 0.026 m above Mean Sea Level at Fremantle
mg/L	measure of salinity, expression of the mass of salts dissolved in one litre of water
Normalised AMRR	Adjustment of AMRR by the addition of the lowest value to all values so as to make all values greater than zero
Open drain	A dual purpose groundwater/surface water drain that is not completely enclosed within levee banks
Radial flow	Groundwater flow towards the wetted perimeter of the drain whereby the flow-lines resemble converging radii (Ritzema 1994)
Rain-fed recharge	Recharge from the percolation of rainfall and runoff to the groundwater system (mm)
Recharge	The addition of water to the groundwater system (mm)
Runoff	The volume or depth of water moved over the land surface (kL or mm)
Salinity (specific)	The concentration of total dissolved salts in water or soil (mg/L)
Salinity (gen)/salinisation	The reduction in the productivity or biodiversity of land or water due to an excess of salts within the environment

Salt export	The removal of salt from the aquifer or soil surface by runoff, groundwater movement or drainage processes
Salt load	Salt transported in flowing or dissolved in standing water (t)
Salt storage	Mass of soluble salt in a unit volume of soil (kg/m^3)
Sediment	Material (soil) that is or has been moved from its site of origin by erosion
Specific yield	(μ) The volume of water released per unit of soil from the drainage of an unconfined aquifer. This is equal to drainable pore space because aquifer compressibility has been ignored
Sodic soils	Soil containing sufficient exchangeable sodium ions to adversely affect soil stability and land use. Sodic soils are subject to dispersion resulting in erosion
Soil	The natural unconsolidated mineral and organic material at the surface of the land
Soil (water) storage	(S_S) Water held in the soil profile above the watertable
Surface water channel	A channel constructed for the purpose of catching and conveying surface water runoff
Tonne	1000 kg mass (t)
Transect (bore)	An alignment of bores used to measure a locus/line of points of the watertable
Transmissivity	The rate at which water is transmitted through an aquifer based on its cross sectional area, hydraulic conductivity and hydraulic gradient
Treatment	A specific set of design and construction criteria applied to sections of drain
Unbounded drain	A single groundwater drain that is subject to groundwater inflow from aquifer discharge. The zone of influence (ZOI) of an unbounded drain is greater than the watertable zone of benefit (WT-ZOB) which in turn is greater than its zone of benefit (ZOB)
Unconfined aquifer	A permeable bed partly filled with groundwater the surface boundary of which is the watertable. The groundwater is in direct contact with the atmosphere through the open pore

	spaces of the overlying soil or rock, the upper boundary is the watertable
Water balance	An equation of all of the inputs and outputs of water for a volume of soil or hydrological area over a given period of time
Waterlogging	The accumulation of excess water in the root zone of the soil
Watertable	Surface of unconfined groundwater at which the pressure is equal to atmospheric pressure
Watertable zone of influence	(WT-ZOI) The perpendicular lateral distance on each side of the drain at which drawdown can or has occurred. The outer limit of the WT-ZOB delineates the extent of the drainage catchment.
Zone of benefit	(ZOB) The perpendicular lateral distance on each side of the drain where drawdown has been sufficient to meet the drainage objectives.
Zone of influence	(ZOI) The perpendicular lateral distance on each side of the drain at which there is potential interaction between the drain and the groundwater system. Groundwater movement from the ZOI into the WT-ZOI is aquifer discharge.

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