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

The use of drains and basins to manage salinity at Morawa

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Salinity and land use impacts series

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by

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Department of Water

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Cover photograph: Mid section of the Morawa drain six months after construction
Photographer: S Dogramaci

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An Appendix CD is in the back 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 drain to a basin reduced but was unable to totally prevent the risk of dryland salinity in a north-eastern Wheatbelt trial funded by the Engineering Evaluation Initiative. While the drain removed large volumes of groundwater it did not sufficiently dewater the land area targeted for salinity recovery.

Deep open drains have been trialled in many catchments in Western Australia and are increasingly seen as useful in lowering watertables and reducing salinity impacts on agricultural land. The Morawa drain and basin project was constructed to assess the practicality of lowering the watertable beneath one part of the landscape with drainage while disposing of and containing the discharge in another. In 2004, eight kilometres of groundwater drain were dug to manage a growing salinity problem in the valley floor of a large north-eastern Wheatbelt catchment at Morawa. The drain discharge was pumped into an evaporation basin to prevent the saline water flowing downstream.

The results of the project challenge the general understanding that the watertable responses to drainage are controlled and limited only by the permeability of the soils and aquifers beneath Wheatbelt valley floors. This understanding has largely come about because, in most cases, only watertable level changes in response to drains were measured and the drains consisted of single channel schemes.

At Morawa, monitoring the watertable alone provided inadequate information on the efficiency of drains. The watertable fluctuations alongside reflected combined responses to recharge, evaporation, lateral flow and drainage. It was impossible to accurately measure the impacts of the drain alone on lowering the watertable.

Only when drain discharges were accurately measured in conjunction with watertable responses could the relationships between the drain and its surrounding aquifer be more clearly understood. Comprehensive monitoring and calculating the water balance revealed that, even though more than 250 000 kL/yr of groundwater were drained, the expected significant fall in the watertable alongside the Morawa drain was not seen.

The findings from Morawa suggest that the single drain layout combined with a highly transmissive underlying aquifer undermined the drain's capacity to lower the surrounding watertable in the face of a seemingly endless groundwater supply. Even with increasing soil permeability it is likely that the combination of groundwater head and the size of the groundwater catchment would only result in increased rates of drain discharge at the expense of lowering the watertable.

Even if the soil and aquifer are sufficiently permeable, the physics of groundwater flow can make the single channel approach to groundwater control for reclaiming saline land ineffective. Preliminary research suggested that the most efficient way of lowering the watertable at Morawa is to construct a drainage scheme with parallel drains spaced at about 370 m apart. The low rainfall and low recharge rates would be expected to improve the

potential for success, but as yet this suggestion remains tested and proven only at Dumbleyung in the southern Wheatbelt.

The underlying transmissive aquifer had a profound effect on basin leakage. Even with pumping from a cut-off drain surrounding the basin, leakage rates became unmanageable and about 80% of the water pumped into the basin was eventually lost to leakage. The leakage was only contained within the property because the area around the basin is very flat and the leaked water evaporated from the shallow watertable.

Analysis showed that there is probably no capacity to hold water above the watertable level at Morawa without high rates of basin leakage. The cost and feasibility of lining such basins are likely to be prohibitive. Basins excavated to below the watertable would end up resembling the saline playa lakes of the region as the salts accumulate within them. Regardless of leakage, farmers with access to areas of saline land, as at Morawa, could probably still safely use these areas for disposing of saline water. Rather than relying on containment by, and evaporative loss from, basins, containment would be by hydraulic gradient and evaporative loss from the resultant high watertable surrounding the basin.

Overall, this report highlights that comprehensive drainage design and assessment need practitioners and operators to investigate and understand the whole groundwater system. The presumed responses of drains and basins should not be viewed in isolation from the surrounding catchment and underlying aquifers.

1 Introduction

The water balance in the Western Australian Wheatbelt has changed since the land was cleared of its deep-rooted perennial vegetation. In some areas after clearing, rain-fed recharge to aquifers has increased up to 100-times. Increased recharge caused groundwater to rise, mainly beneath the broad valley floors in the poorly drained landscapes. Rising groundwater usually carries dissolved salts, mobilised from their historical accumulation and storage within the underlying soils. The saline land area affected by mobilised salts is expected to expand to over three million hectares by 2020 (Nulsen 1998).

In the 1970s, landholders began to advocate and dig groundwater drains to increase discharge and manage rising watertables, waterlogging and associated salinity (Coles et al. 1999). Implemented on a farm-scale, drains 1–3 metres deep were constructed to intercept groundwater; they became commonly known as ‘deep drains’.

Thirty years on, and the total length of earthworks for salinity control in the Wheatbelt exceeds 10 000 km (ABS 2003) of which at least 5000 km are deep drains. Yet, there is mostly only anecdotal evidence on the impacts and benefits of deep drains. Much of this evidence for reclamation of saline land by drainage appears to be related to seasonal variability. Discharges into existing natural waterways can still cause concern within the community, particularly those downstream, about the volumes and quality of the waters involved.

Evaporation and detention basins have been suggested as solutions to manage and dispose of saline drainage water on-farm, minimising damage to downstream properties, waterways and environments. However, little is known about the integration of basins into Wheatbelt drainage projects and their behaviour within the Wheatbelt landscape.

In 2002, the Government of Western Australia initiated the Engineering Evaluation Initiative (EEI) through the predecessor of the Department of Water. The EEI was established to evaluate various engineering solutions for salinity without further damaging the environment. It was a $4 million priority project under the National Action Plan for Salinity and Water Quality. The EEI consisted of three programs (Dogramaci & Degens 2003):

1. Improved siting and design of engineering options to maximise performance at the farm scale
2. Safe disposal of discharge waters
3. Implementation of options within a planned regional drainage context

The main scope of EEI was a focus on increasing understanding of the appropriate use of engineering options to manage dryland salinity for economic, social and environmental benefits (Dogramaci & Degens 2003). As part of the EEI, the Department of Water conducted cooperative salinity reclamation trials with landholders across the Wheatbelt. This north-eastern Wheatbelt trial investigated a farm-scale deep drainage and evaporation basin scheme.

Measurement of the groundwater conditions began six months before and ended two years after construction of the drain and basin at Morawa. Drain discharge and basin performance were also measured for two years after construction.

The principal aim of the Morawa EEI project was to evaluate the effectiveness of a deep open drain on lowering the watertable, and the usefulness of a local-scale evaporation basin to dispose of the discharge from the drainage scheme. To a large degree this aim could only be achieved with a monitoring program established to identify and measure as well as possible all aspects of the water balance that affected the scheme.

The specific study components were:

- Define the aquifer characteristics at the site.
- Investigate the effectiveness of the drain in lowering the watertable.
- Better understand the drivers of watertable response to the drain.
- Evaluate the performance of the farm-scale evaporation basin.
- Calculate the water balance at project scale.

1.1 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 to understand 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 from the data analysis and interpretation

The CD is provided in the back of the printed report. Online readers can request the CD from the Department of Water.

1.2 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 many significant figures to enable the reader to follow the evolution of, and interrelationships between results and to show results that might have very low values.

It is acknowledged that, even without gross errors, individual measurements have inherent inaccuracies associated with instrument error and field techniques. Suggested levels of systematic error for the variables measured for the Morawa project and used in this report are:

- Groundwater level measurements: +- 0.05 m
- Pumped main-drain discharge: +- 10%
- Mid-drain discharge (by stage): +- 20%
- Basin water levels: +- 0.02 m
- In situ salinity: +- 5%
- In-situ pH: +- 0.5 units
- Continuous salinity: +- 10%
- Rainfall: +- 5%
- Pan evaporation: +- 10%

The systems approach used here should have limited reaching erroneous conclusions from the measurements and uses data trends and relationships, rather than focusing on individual sets of results. However, it is acknowledged that significant extrapolation of some results has been needed given the temporal and spatial distribution of some elements of the monitoring program. For example, water level change across the 597 ha drainage site was measured with an average of one bore per 33 ha.

This report has deliberately steered away from using complex hydrological models in this initial assessment of the Morawa drainage project so that the main variables that affected the performance of the drains and evaporation basins at Morawa are identified and clarified. This more simplistic approach is also likely to be replicable, which may not be the case for complex models where the variables underpinning assumptions and data relationships may be hidden.

2 Drain and basin site

Fairway was selected as an EEI study site with the support of the Morawa Farm Improvement Group, of which the property owners Rod and Lorraine Madden were also members. Its shallow indurated cemented subsoils offered an opportunity to construct and monitor drains in highly permeable aquifer conditions.

The expectation was that the cemented subsoils would provide an open fabric with preferred pathways that allowed groundwater to migrate towards the drain from beneath the wide valley floor. The expected result was an improved capacity to lower the watertable at greater distances from the drain than previously experienced in the plastic clays and loams elsewhere in the Wheatbelt.

Fairway is situated approximately 28 km north of the Morawa townsite along the Morawa–Yalgoo and Madden Roads (Fig. 1). The project affects 3564 ha of the property on Victoria Locations 8378, 6953, 8347 and Lot 2 on Plan 7135. The evaporation basin and discharge end of the drain are at (MGA 50) 404 500 mE, 6 794 700 mN.

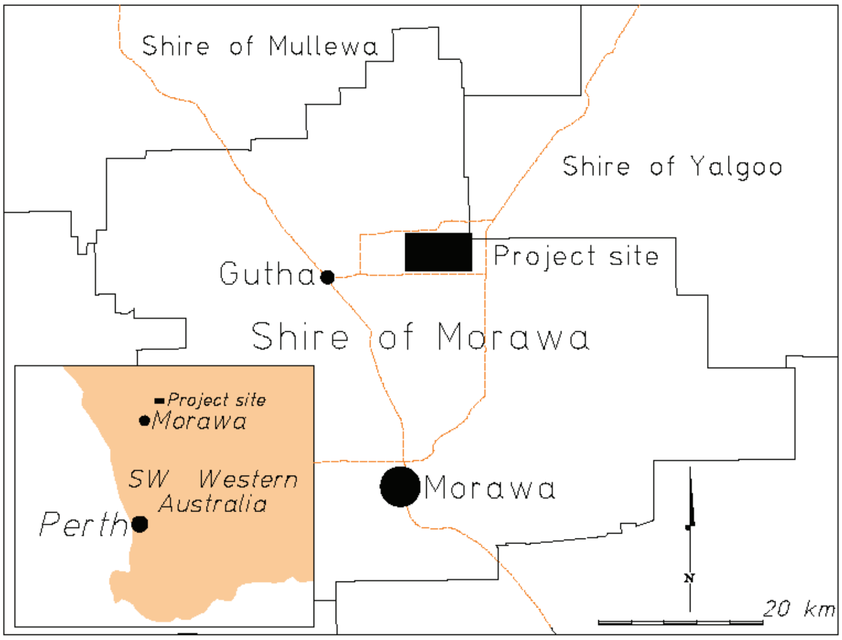

Figure 1 Location of the Morawa project site

2.1 Climate

The site has a Mediterranean climate, with warm to hot summers and mild, moderately wet winters. Bureau of Meteorology records show that Morawa had experienced an average of 20.2 days per year with maximum temperatures at or above 40 °C for the period 1911–2005. Maximum temperatures rose to or above 30 °C for an average of 130 days per year and minimum temperatures fell to 0 °C or less on average about 1 day in 3 years.

The winter season, beginning in May and ending in September, usually accounts for about than two-thirds or 65% of the annual rainfall. Winter rainfall is usually associated with cold frontal systems originating from the south-west. These rainfall patterns affect large areas and are generally of low intensity and longer duration than summer events.

Rainfall during summer is usually extreme and highly localised – the result of heavy thunderstorms that are sometimes cyclonic in origin. December to early May rainfall is often the product of moist air being drawn south from the north-west. Summer rainfall is unreliable and not uncommonly totally absent.

An automated rain gauge and pan evaporimeter (Station number 508042) were operated at the evaporation basin from November 2004 to January 2007 (Fig. 2). Over the 774 days of operation, 420.8 mm of rainfall and 5826 mm of A pan evaporation were recorded. Rainfall included three significant events over 20 mm/d (all in late summer) with the highest recorded daily value of 23.4 mm/d measured during March 2006 (Fig. 2). Potential evaporation is exacerbated in spring and summer by dry easterly winds blowing from inland areas.

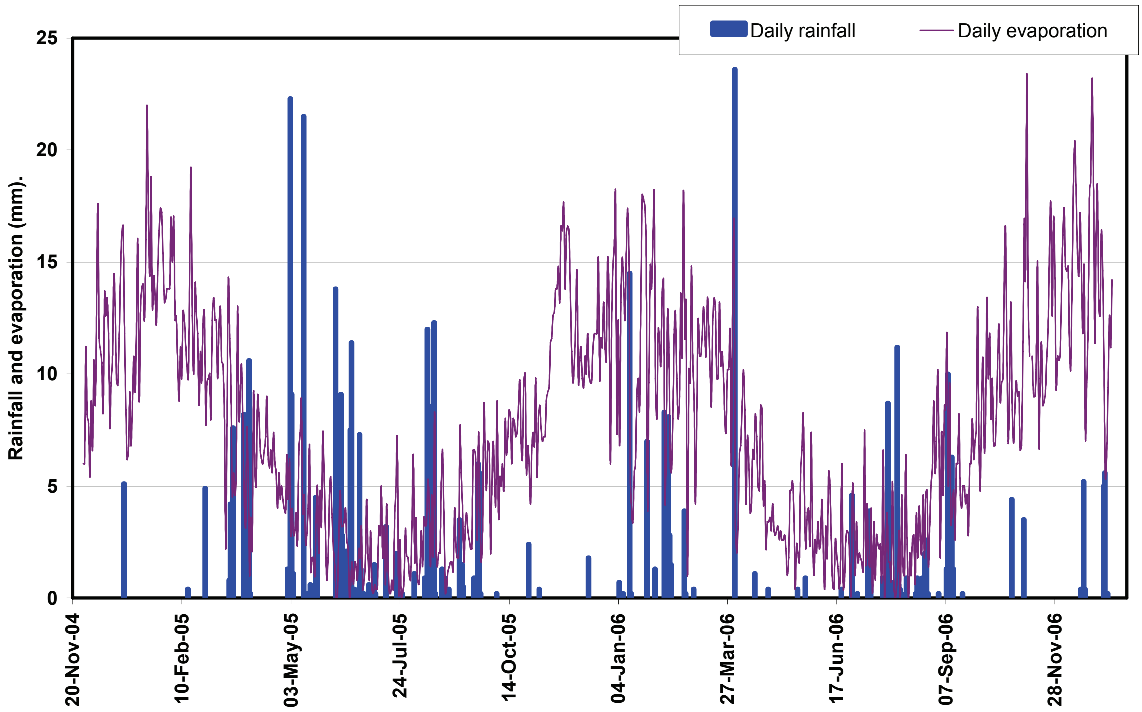

Figure 2 Daily rainfall and evaporation from the automated rain gauge and pan evaporimeter (station 508042)

The 2006 site rainfall of 155 mm was less than half the 1970–2007 average of 322 mm/yr for Morawa (BoM 1998). Morawa's 1970–2007 range was 172.5–519.8 mm/yr and its average (322 mm/yr) was just higher than its long-term (1889–2007) average annual rainfall of 320 mm, range 118–580 mm/yr and median 328 mm/yr.

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

An analysis of AMRR for Morawa 1970 to 2007 (Appendix CD 2.1a) shows a general decline since 2000 (Fig. 3). The average monthly rainfall for 1970–2007 was 26.8 mm. During the low summer and higher winter rainfall seasons for Morawa, the AMRR should be in decline from around October and starting to rise again in May.

Large increases in AMRR outside the seasonal trend are mostly driven by summer and early winter rainfall events as discussed above. These rainfall events are often in excess of 30 mm/d and occurred almost annually until 2001. Over the period 1970–2001 there were 11 rainfall events in excess of 60 mm/d and 37 events of 30 mm/d or more. Since 2001 there were only two rainfall events in excess of 30 mm/d; both during 2006.

Because watertable levels appear to correlate with both AMRR trends and individual larger rainfall events, the absence of larger events produced a falling AMRR trend which could possibly be reflected in falling groundwater 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 a drain response could be attributed to this natural groundwater decline.

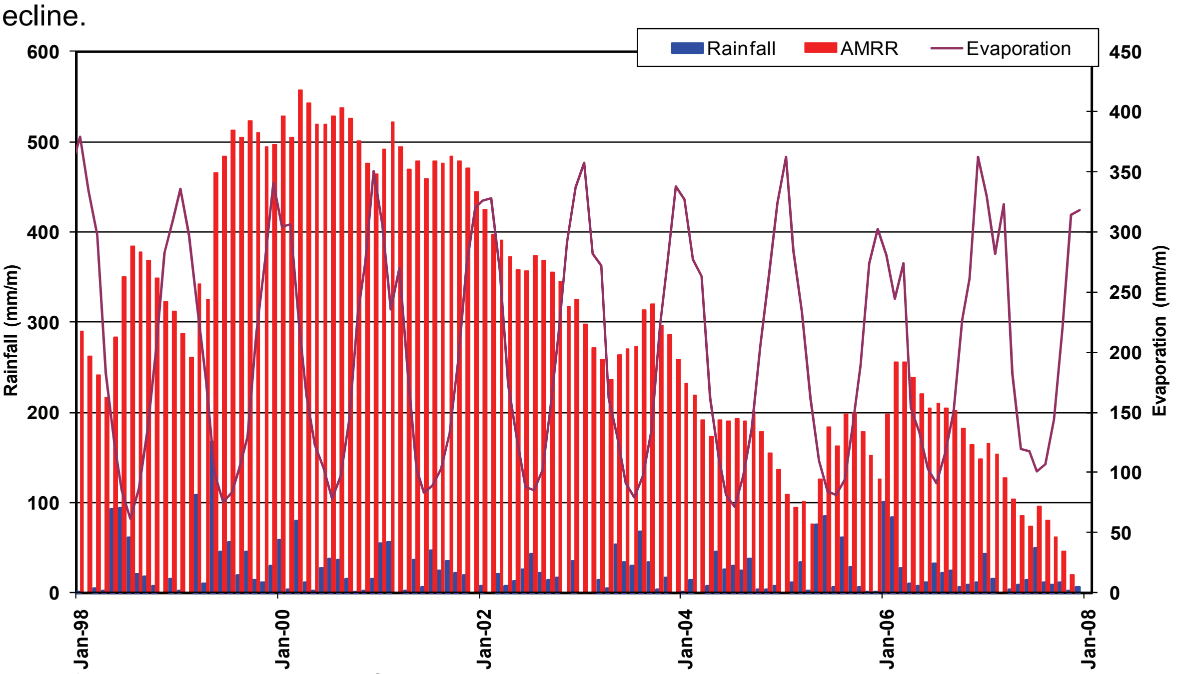

Figure 3 Morawa monthly rainfall, AMRR and evaporation

It is not known if the pre-drainage AMRR trend for the project site was similar to the decreasing trend of Morawa given the 27.5 km distance between them. There was a significant difference between the rainfall in these gauges during the corresponding period of measurement (December 2004–06). The average monthly rainfall for Morawa (2005–06) was 27 mm compared to 16 mm for the project site. This was largely due to the 49 and 59 mm/d rainfall events experienced at Morawa in January and February 2006 that are almost totally absent from the rainfall recorded at the project site. This discrepancy reflects the localised nature of summer thunderstorms.

With an average monthly rainfall of only 16 mm for the period of analysis, the decrease followed by only small increases in AMRR measured at the project site during April–September 2006 reflects that year's very dry winter conditions (Fig. 4).

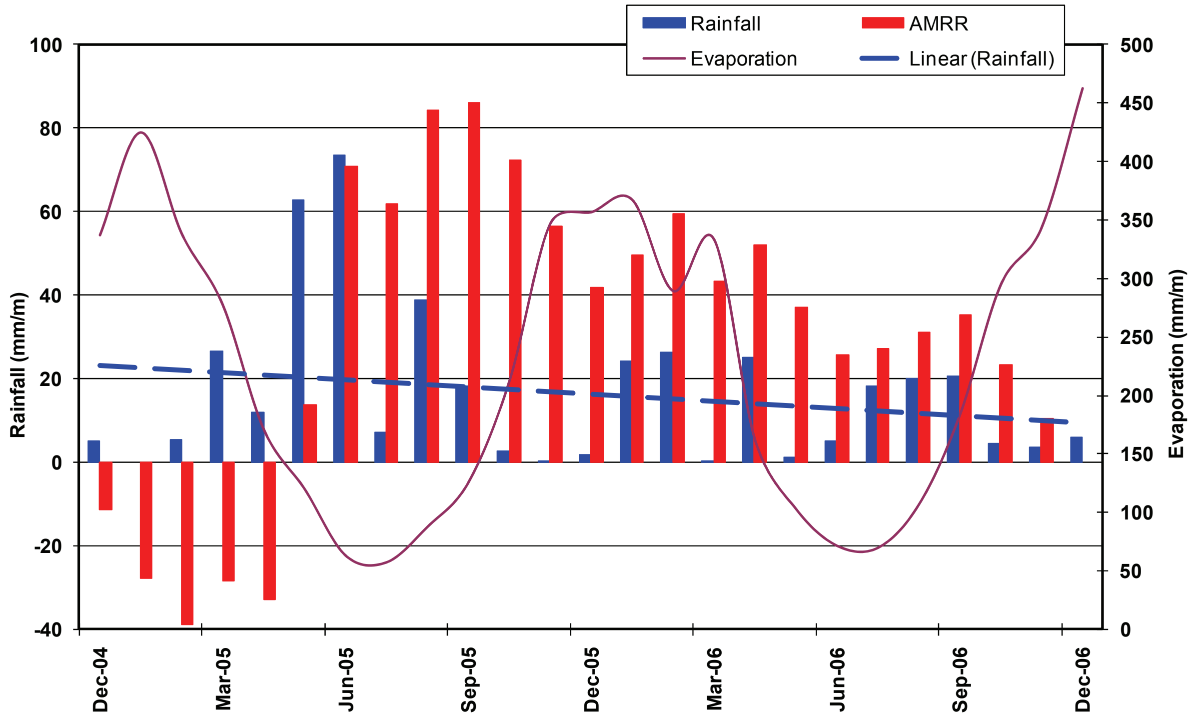

Figure 4 Monthly rainfall and linear trend, AMRR, and evaporation from the project site

Morawa AMRR was normalised for the period June 2004–December 2006, and adjusted to provide values for the corresponding pre-drainage watertable measurements (June–November 2004). Normalising the Morawa and site AMRR so that their minimum values equated to zero highlighted the similarity of their trends (Appendix CD 2.1b). A significant deviation in the trends only occurred in response to the 49 and 59 mm/d rainfall events at Morawa but not at the site in early 2006.

The normalised Morawa AMRR values for June–December 2004 were then downscaled by 39 mm to equate the December 04 value with that of the site. The resultant AMRR trend is a combination of normalised and adjusted Morawa values from June–November 2004 and normalised site values from December 2004–06. This is further referred to as the AMRR trend used in this report.

2.2 Land use

In the Shire of Morawa, 80.6% of its 342 000 ha was cleared for agriculture by 2002 (Shepherd et al. 2002). The current agricultural land use is predominantly dryland cropping of cereals and legumes during the winter and, dependent on seasonal influences and livestock prices, grazing (mostly sheep) on crop residues, introduced pastures and perennial shrubs.

Nearly 96% of Fairway was cleared of native vegetation by the 1960s, but 948 ha has since been colonised with halophytes on the valley floors and associated areas as they became progressively salt-affected after clearing. The last area of about 40 ha of salt-affected land was excluded from cropping around 1996. Halophyte areas of mostly the volunteer shrubs blue bush (*Maireana* spp.) and saltbush (*Atriplex* spp.) (Fig. 5) presently cover approximately 1009 ha or 28.3% of the property.

The salt-affected and halophyte-covered land provides limited economic return beyond that derived from the opportunistic grazing of livestock. The Morawa EEI drain and evaporation basin are surrounded almost entirely by this type of land use.

Figure 5 Saltbush in the drainage and basin site

2.3 Topography and natural drainage

Fairway is characterised by low relief bordering the broad central valley floor selected for this drainage and basin site. The valley enters the northern boundary of the property at an elevation of 289 m AHD, deviates SW and exits the western boundary at 283 m AHD.

The valley floor width narrows from about 2000 m to 700 m at its mid-point through the property. This constriction has side slopes that rise at average gradients of 1.25% to hilltops with elevations at about 310 m AHD. At the constriction point, the valley floor gradient is 0.11% compared with the average 0.04% both up and downgradient. Two playas, about 80 m and 120 m in radius, are about 1500 m upgradient from the constriction and a third is just beyond the western property boundary.

The landscape to the south and east is gently undulating except for some isolated hills and ranges, some topped by mesas at about 390 m AHD elevation (Fig. 6).

The surface water catchment upstream from the western property boundary is 1080 km^2, reaching a maximum elevation of 392 m, some 60 km to the NNE. As the catchment headwaters rise within the rangelands, little of the native vegetation has been cleared for agriculture. A significant tributary crosses the eastern boundary of the property and has a catchment of 61.5 km^2 of the 1080 km^2.

Discharge from the catchment and property flows westward, entering the salt lake chain of the Yarra Yarra palaeo-river system. This system is a non-contributing part of the 66 093 km^2 Moore River catchment (Mayer et al. 2005). Being an internally drained system the salinity of flowing surface water within the Yarra Yarra and most of its major cleared tributaries ranges from saline to highly saline (7000–35 000 mg/L). The high evaporation rate during summer usually forms brine (> 35 000 mg/L) from any natural free-standing water.

Figure 6 Flat valley floor to hilltops

2.4 Soils

Alkaline, red, deep loamy duplex profiles dominate the soils in the valley floor (Schoknecht 1997). These typically consist of red and red-brown clayey/loam topsoils 0.3–1.8 m thick (Fig. 7). Only minor areas of these topsoils that are located close to the valley footslopes are hard-setting. Some surface sealing of saline topsoils also follows significant rainfall, reducing infiltration.

Topsoils abruptly give way to cemented calcareous subsoils, sometimes interspersed with red, grey-brown and grey clay through to sandy clay. Some subsoil is fractured platey bedded layers from about 0.5 m to below the 3 m test-pit depth. Silcrete layers are often identified within and beneath the calcareous subsoils. The depth to subsoil generally decreases down the valley to the SW from about 1.3 m near the head of the drain to less than 0.4 m at the basin (Section 3 Fig. 11).

Cemented subsoil did not limit drilling which returned a sample similar to sandy loam in texture (Dogramaci 2004). Excavation with a backhoe and sometimes an excavator proved more difficult, particularly where silcrete was present, until the material was penetrated and could be lifted and fractured.

Shallow silcrete was mostly encountered when excavating the drain through the constricted section of valley floor (Section 2.3). Silcrete at the outer end of the SE drain tributary was difficult to penetrate or lift, so construction of this section of drain was abandoned (Fig. 11). The soil here possibly has the only silcrete hardpan that sometimes restricted infiltration, causing a perched watertable.

Figure 7 Soil profile typical of the Morawa drain showing loamy topsoil over cemented subsoil

2.5 Geology and hydrogeology

The property is within the Archaean Yilgarn Craton, west Yilgarn tectonic terrane (Nulsen 1998). The landscape comprises broad valleys of Cenozoic and Quaternary sediments. These overlie the crystalline basement: granitic and gneissic rock intruded by dolerite dykes. Some basement rock outcrops occur on the valley flanks within the property and surrounding landscape.

Regolith varies in depth and composition, generally consisting of one or more of:

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

McGowan et al. (1987) contoured groundwater salinities in the order of 5000 to 20 000 mg/L within the region, and spot salinities that range between more than 36 000 mg/L (seawater) within the valley floors to less than 275 mg/L (very fresh) in hillside sandplain seeps (Nulsen 1998).

Forty-nine bores were drilled in the valley floor to determine the hydrogeology (Dogramaci 2004) and to monitor groundwater levels and quality (Section 3.3). The drilling (Appendix CD 2.5a) indicates that the weathered profile consists of three hydrogeological layers (Fig. 8). The deepest layer developed above the basement by the formation of angular medium to coarse grains in an approximately 5 m thick layer from the fragmental disintegration of basement rocks rich in quartz and feldspar.

Above this layer lies a 15–20 m thick clay-rich layer. This second layer appeared semi-permeable, impeding the upward movement of groundwater from the deepest layer to the surficial sediments that cover much of the valley flats, although no significant difference between groundwater head and watertable of the deep and surficial aquifers was detected.

Quaternary surficial sediments of medium to coarse-grained quartz sands and clays overlie the in-situ weathered profile. Surficial deposits with alluvial, colluvial and aeolian origins have higher hydraulic conductivities (Section 2.5) than the clay-rich layer. Silcrete 1 m thick was encountered in most of the bores 2–4 m from the surface, returning coarse angular sand during the drilling. The silcrete layer within 4 m of the surface might indicate a historically high watertable. Figure 8 provides a three-dimensional view of the topography, bores and aquifer beneath the drainage site.

The hydraulic conductivity of the deep layer varied because of the mineralogy and location in the landscape. The airlifting of four deep bores indicated a hydraulic conductivity of about 1.2 m/d. Groundwater movement is relatively fast in this layer. For the intermediate (8–12 m deep) bores airlift discharge rates indicated a hydraulic conductivity of approximately 0.5 m/d. Airlifting of the shallow bores also indicated a hydraulic conductivity of about 1 m/d (Dogramaci 2004). The estimation of hydraulic conductivity with airlift discharge is only indicative. Drain discharge, groundwater trend analysis and groundwater modelling provide better insight into the groundwater dynamics and the hydraulic properties of the aquifers.

The hydraulic conductivity (K_{sat}) in six of the shallow bores was measured directly using the rising slug test method (Freeze & Cherry 1979). The bores measured included the four monitoring bores 50 m from the drain, the comparison bore 032 and the basin monitoring bore 027. The average hydraulic conductivity was 0.34 m/d, with a standard deviation (SD) of 0.28 m/d and range 0.10–0.72 m/d. The high standard deviation was the result of the higher values for the comparison and basin bores (0.66 m/d and 0.72 m/d respectively). In the absence of these values the average for the transect drain monitoring bores was 0.16 m/d with a SD of 0.06 m/d.

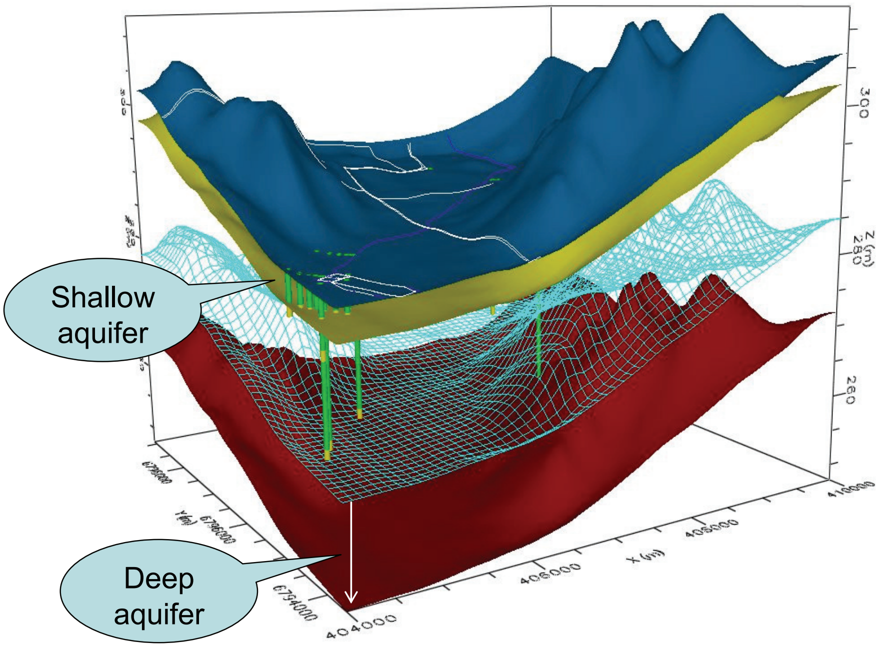

Figure 8 Project site, bore transects, evaporation basin and aquifers

2.6 Watertables and salinity

Groundwater levels within the valley floor had risen so much that evaporation from the soil surface was leaving salt residues in plant root zones. Anecdotal evidence (R Madden, pers. com.) suggests the first expressions of salt appeared after floods in the 1960s. Since then, the area of salt-affected land within the valley floor has increased with the rising saline watertable.

The high permeability of the sandplain soils at the footslopes produces saline discharge and severe scalding by capillary action from the shallow watertable. This could be seen as indicating that hillside aquifer discharge was a contributor to groundwater recharge in the valleys and the likelihood of elevated groundwater levels beneath the valley flanks.

Drilling provided groundwater levels, and salinity and pH measurements beneath the valley floor (Section 4) but there are no historical groundwater data or any indication of the depth to watertable beneath the surrounding elevated land. Groundwater studies (Nulsen 1998; Speed & Kendle 2008) differ on previous likely watertable change beneath the valley floor. Possibly due to recent dry seasons watertables have fallen on the valley flanks but not beneath the valley floors.

All the monitoring bores indicated a fairly static watertable at not more than 1.0 m below ground level (bgl) across the valley floor. As the watertable fairly closely followed the contours of the surface topography, it is thought that watertable height had remained fairly static over the recent past. The understanding is that the watertable would have been at this approximate level, at equilibrium between in-situ recharge, recharge from the valley flanks and discharge by evaporation or transpiration from the soil surface.

The groundwater salinity increased overall from the upper to the lower valley floor and from the footslopes towards the centre of the valley (Section 4.2). The average shallow salinity in the bores (MW001–006) at the upper valley end of the property was 15 890 mg/L in 2004. The average for the deeper groundwater salinity was about 10 000 mg/L higher than this. In the lower portion of the valley average shallow groundwater salinity was 38 203 mg/L (seawater has salinity around 35 000 mg/L), with a deep groundwater salinity of 40 290 mg/L.

These groundwater salinities are representative of a Wheatbelt valley floor environment and the quality of the groundwater that might be drained (Chandler & Coles 2003). With a depth to basement of 40 m and suggested aquifer porosity of 30%, 40 290 mg/L of salt in solution equates with salt storage of 4835 t/ha beneath the valley floor.

3 Morawa project design and methods

3.1 Drain design and construction

Construction of the drain initially took about 57 days from December 2004 to February 2005. Due to hard ground conditions (Section 2.4) the planned 8885 m of drain fell short by 893 m. The completed 7992 m of drain included 690 m surrounding three sides of the evaporation basin (to capture leakage) colloquially referred to as the 'cut-off' drain, and 7302 m left as the main drain.

Then in September 2005 a 'rock saw' began to excavate north of the 893 m shortfall and reconnect from north of the planned route and join with the main drain. This was done in an attempt to improve drainage in this vicinity, before approaching and continuing along the alignment of the originally abandoned drain section (Fig. 11). Near the head of the existing drain, ground conditions proved too hard and excavation stopped when costs approached the equivalent of $80 000 /km. An additional 950 m of vertically-sided 0.6 m wide drain (Fig. 9) was excavated using the rock saw, taking the total length of groundwater drain to 8942 m.

The original section of drain was dug using a 52 tonne bucket excavator proceeding upstream along the alignment. The bucket had a trapezoidal profile to enable it to cut 1:0.5 (V:H) batter slopes. The original specifications were 2.5 m depth, 1.1 m width bottom and 1:0.5 batter slopes, forming a channel with a cross-section of 5.8 m^2. The spoil was formed into continuous 1.1 m high enclosing levee banks along each side of the channel. As the drain discharge was to be pumped into a basin, the width of the berms was reduced to 2 m to limit surface water capture from rainfall. This kept the total width of the catchment area of the drain structure between the levee banks to about 12 m.

The drain was as close as possible to the lowest alignment along the valley floor while still leaving floodway capacity for significant runoff events from the 1080 km^2 upstream catchment. Towards its discharge end the drain was diverted SW, away from the main natural drainage line, further isolating the basin from potential flooding events. Designed to convey almost all potential surface water flow along the northern side of the drain, floodwater crossings over the channel were unnecessary and only one essential culvert crossing was required under Holders Road (Fig. 11).

Pre-construction site investigations included preliminary and final elevation and position surveys for the drain alignment and surrounding features, and a geotechnical investigation with open pits. The average gradient of the floor of the 7 km main drain was 0.08%. Its starting elevation was 289.37 m AHD at the northern boundary of the property, and terminated at 283.00 m AHD at the NE corner of the basin.

Eight pits dug using a backhoe to investigate digging conditions and soil permeability mostly encountered hard conditions at or below the proposed drain depth and watertable. This limited the assessment of soil permeability to only two pits. These yielded hydraulic conductivities (K_{sat}) of 0.11 and 1.10 m/d with their mean of 0.60 m/d adopted for drain and basin design. The hydraulic conductivities had not been estimated from the bore slug tests at this time (Section 2.5) as the bores were not yet drilled.

The Hoodghoudt Steady State Equation (Ritzema 1994) was used pre-drain to estimate discharge from the drain specifications; K_{sat} value (0.60 m/d); the recharge or drainage coefficient (10% of average annual rainfall = 32 mm/yr); the thickness of the aquifer beneath the drain (10 m); and the starting watertable (1.14 m bgl). For the planned 8885 m of drain the estimated annual steady-state discharge was 217 905 kL and the zone of influence (ZOI) was 365 m each side of the drain.

The aquifer thickness of 10 m was used in the performance estimations following a review of other drainage evaluations conducted in Buntine, to the south (Speed & Simons 1992). Results from Buntine suggested drainage effects would be limited to horizontal flows within the surficial aquifer and towards the drain, irrespective of the underlying aquifer characteristics. This conclusion also led to abandoning the comparatively high estimated steady-state discharge and calculated zone of influence, the results of which were more appropriate indications of the Morawa drain's performance.

Figure 9 The rock-sawn drain section

3.2 Basin design and construction

In place of the calculated drain discharge figures above, discharge into the evaporation basin (basin) was calculated based on measured and estimated values from other Wheatbelt drains. This resulted in constructing an under-designed basin with a surface area of 5.12 ha and design storage volume of 48 000 kL. The principal source of water loss from the basin was intended to be evaporation from its surface, i.e. an evaporation basin. Provision was made for basin expansion and for regulating the discharge from the drain by the pumping rate.

The original basin had two roughly equal square cells linked by two 150 mm diameter pipes through the dividing wall near the natural ground level. The basin was constructed with a small bulldozer pushing up 1.5 m high walls with a 1:5 inside batter and 4 m wide top, from inside each of the two cells. This left the walls and their borrow areas surrounding central areas of unaffected flat land that would inundate as the basin filled to above ground level (Fig. 10). The borrow areas for the walls were kept at 0.75 m in depth or less to avoid excavating into the watertable beneath the site and bogging the bulldozer.

'Track rolling' of the basin walls during construction was the only compaction – there being no advantage elsewhere in the cells, due to the low soil clay content, the presence of cemented hardpans and the shallow saline watertable. The rationale in the site selection was that basin leakage would be controlled by the cut-off drain, aquifer conditions beneath the site and the hydraulic gradients extending away.

Figure 10 Evaporation basin wall and partially filled borrow area

Basin leakage, together with the shallow encompassing watertable, was controlled by the 690 m of 2.5 m deep cut-off drain that connected to the main drain at the NE corner at the same invert. Their combined discharge was lifted into the basin from a 3 m deep sump in the lowest (NW) corner of the basin using a 150 mm engine-driven centrifugal pump with a rated capacity of 5200 kL/d. Until early 2006 the pump was driven by a tractor started manually by the landholder, after which it was coupled to a diesel engine that started automatically, controlled by the depth of water in the sump.

By 15 June 2005 the original basin had filled and was extended to allow pumping from the drainage scheme to continue. The extension involved constructing a 26.6 ha single-cell contingency basin adjoining the southern side of the existing basin. The enclosing wall of the new basin was constructed with a bucket excavator to 1 m height and base width around 3.0 m. Like the original basin, the earth was borrowed from the inside near the base of the wall, leaving the remainder of the enclosed area as undisturbed ground. A 1.2 m deep cut-off drain was constructed along the southern and western sides of the basin (Fig. 11) to divert shallow seepage to the head of the existing western cut-off drain. The new basin was connected to the eastern cell of the original basin with a 150 mm PVC pipe with a screw cap passed through the south wall at the original ground level.

At this time the pump station was also relocated to draw water directly from the end of the main drain at the NE corner of the basin, and the main drain disconnected from the cut-off drain. Pumping from the cut-off drain ceased at this time which separated the 4062 kL temporary storage of the cut-off drain and sump from the end of the main drain. However, 6679 kL of temporary storage remained in the main drain providing an adequate reservoir from which to pump discharge into the basin.

3.3 Monitoring methods and data availability

There was a strong emphasis on measuring watertable changes associated with discharge from the drain, inflow to the basin and all possible sources of groundwater loss. This was so that the effects of these salinity engineering works could be quantified and sound correlations or 'cause and effect' relationships identified.

The monitoring program had to be responsive to the adaptive management of the drainage scheme (Section 3.2). All components of the program and related management factors are itemised below. The monitoring works locations are shown on Figure 11 and the detailed project plans in Appendix CD 3.0.

Rainfall and evaporation

An automated (pluvio) rain gauge and A-pan evaporimeter (station 508042) recorded from 26 November 2004 to 10 January 2007 at 5 minute and 20 minute intervals respectively. Both were located on top of the NW corner of the evaporation basin wall at coordinates (MGA 50) 404 260 mE, 6 794 520 mN.

Watertables

Thirty-nine bores (MW001–032) were drilled across the valley floor (Section 2.5) prior to drain construction to determine pre- and post-drain water levels (Fig. 11). The water levels were measured fortnightly to monthly from 13 May 2004 to 2 March 2007. Automated water level recorders operated at six-hourly intervals in all of bores from 13 November 2004 to 19 February 2005 during initial drain construction. They were re-installed in bores MW002, 008, 014, 020, 027 and 032 to capture ongoing watertable trends from 1 December 2005 to

2 March 2007 when the in-situ measurement frequency was reduced from fortnightly to monthly.

In-situ salinity and pH in all bores were measured approximately three monthly. A full suite of major ions, metals, N and P were laboratory analysed (Appendix CD 3.3) from samples taken from the deep and shallow bores MW001, 007, 013 and 019 and comparison bores MW030 and 032. The samples were collected at the end of the summer and winter 2005 and 2006.

Twenty-eight of the bores, arranged in four transects, were used to measure the watertable responses to different sections of the drain. Transect 1 contained bores 001–006, transect 2, 007–012, transect 3, 013–018 and transect 4 up to 024. Transects were arranged perpendicular to the drain with the closest deep and shallow nested bores 20 m from the channel. The other shallow bores were 50, 100, 175, 275 and 400 m from the channel.

Ten additional bores (MW033–042), measured together with the 39 bores, provided additional groundwater data from 22 November 2005 to 23 November 2006. Most were drilled on the property to the west of the project site (Victoria Location 8378) to monitor watertables downstream of the basin.

Bores are referred to by number only in the rest of the report; for example, MW001–042 as bores 001–042. Intermediate and deep nested bores sharing the same number as their shallow counterpart are identified with the suffix ‘i’ for intermediate and ‘d’ for deep, corresponding to the drilling log (Appendix CD 2.5a). Nested bores are less than three metres apart.

Drain discharge

Pump station

Discharge was measured at two points along the drain. The more comprehensive of these monitoring stations was the end-of-drain pump station that lifted drain discharge into the evaporation basin. Station 618605 (605) operated from the start of drain construction on 22 December 2004 to the project’s end on 10 January 2007. The discharge volume was measured by a flow meter fitted with an EC probe and a data logger that recorded pumped volume and salinity at five minute intervals.

From the start of pumping to 15 June 2005 the pump station was located at the NW corner of the basin and was lifting drain discharge and captured basin leakage into the western cell of the basin. When the cut-off drain was disconnected (Section 3.2) it was relocated to the NE corner of the basin at (MGA 50) 404 529 mE, 6 794 704 mN and lifted only drain discharge into the basin’s eastern cell.

Drain discharge was laboratory analysed at the same time as the groundwater and for the same suite of major ions, metals, N and P.

Sump water level

The water level in the drain sump or end of the drain station (606) was measured hourly by a capacitance water level sensor. These measurements were used to calculate the volume of water stored at the inundated end of the drain. The volume of water was calculated from the depth-to-volume relationship established from a bathymetric survey of the sump, cut-off and lower main drains (Appendix CD 3.1).

From 22 December 2004 to 10 June 2005 the water level sensor was located in the drain sump. After the main drain was disconnected from the cut-off drain, the sensor was relocated next to the pumping station intake in the end of the main drain and operated from 14 March 2006 to 8 January 2007.

Mid-drain station

A 'V' notch weir, fitted to the inlet end of the culvert at Holders Road (MGA 50, 406 130 mE, 6 795 700 mN), provided drain-flow measurements from 29 April to 16 July 2005. This station and sample site 025 was later relocated up the drain (MGA 50, 406 952 mE, 6 798 617 mN) to avoid flooding caused by tail-water inundation from the pump station (Fig. 11). It operated at this new location from 19 January 2006 to 10 January 2007.

In-situ salinity and pH of the drain discharge were measured at the same intervals as groundwater. The in-situ measurement site was relocated with the mid-drain station away from Holders Road on 16 July 2005. Before relocation, 5335 m of drain contributed to this station and, after relocation, the contributing drain length was reduced to 4134 m until the additional 950 m of drain was constructed in September 2005.

Evaporation basin

Water levels

The water level in the evaporation basin western cell was measured hourly by a water level sensor. The measurements were used in conjunction with depth-to-volume and depth-to-surface-area rating tables (Appendix CD 3.1) developed from a basin bathymetric survey. This relates changes in the depth of water to changes in the basin storage volume and surface area.

The water level sensor station 607 operated from 22 December 2004 to 9 January 2007 in the western basin cell.

Water quality

In-situ salinity and pH of the basin water were measured at the same time as groundwater, giving records from 3 February 2005 to 6 December 2006. The basin water was laboratory analysed together with the bore and drain discharge program. All measurements and sampling were done in the SE corner of the western basin cell at sample site 607 (MGA 50, 404 460 mE, 6 794 457 mN).

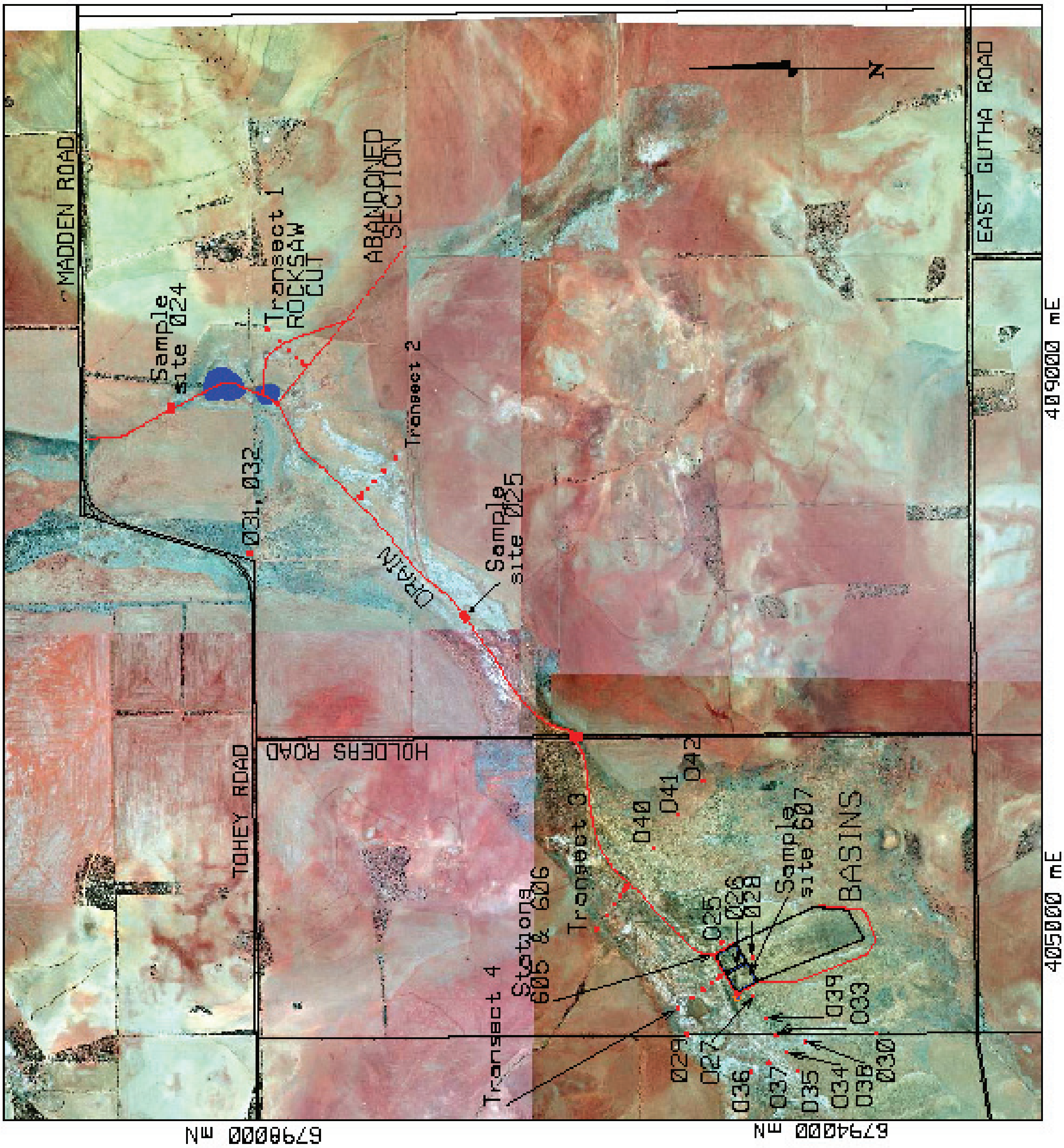

Figure 11 Morawa drain, basins and monitoring sites

4 Results

4.1 Watertable responses to the drain

Comparison bores

Comparison bores were sited to provide background measurements of natural groundwater variability for comparison with groundwater measurements from the project monitoring bores. The spatial variability across the site means that natural processes affect local watertables and groundwater levels differently. Of particular significance might be the relative recharge–discharge relationships between the shallow and deeper aquifers at any given location, and their effects on watertables. The term ‘comparison’ rather than ‘control’ bores is used here to reflect that responses between these and the project bores may consist of similar or different trends rather than changes in absolute values.

Bores 030 to 032 are far enough from the drains and evaporation basin to be unaffected by any groundwater change that they might cause. Bore 029 was apparently close enough to the drain and basin to be affected. So, even though 480 m from the lower section of drain, its levels are not shown or used for comparison.

Bores 031, 031i and 032 are within 20 m of each other and 820 m from the nearest drain. Their hydrographs (Fig. 12) have been reduced to the same local datum (LHD) to assist with trend comparison. There is no apparent difference in the watertable and groundwater trends or levels of these bores beyond that caused by varying measurement frequency.

Bore 030 is 930 m SSW of the original evaporation basin and 660 m from the contingency basin. Its surrounding landscape has a far greater area of flat, salt-affected land than that of the three bores above. In December 2004 the 030 groundwater level was 1.5 m bgl compared to 031’s 1.8 m. The shallower watertable and flatter, more poorly drained environment for 030 may partially explain its more rapid upward response to the onset of rain-fed recharge in June 2005.

The water levels in all of the comparison bores broadly reflect the AMRR trends (Fig. 12). The closest trend relationship is apparent between 030 and AMRR, evident in the corresponding response between the two during October–November 2006 but less evident in the hydrographs of the other bores. The two early May 2005 rainfall events in excess of 20 mm followed by consistent June 2005 rainfall (Fig. 2) caused the most substantial increase in AMRR and corresponding water level rises of over 0.6 m.

Of most interest though is the absence of water level decline in response to a reducing AMRR trend from October 2004–April 2005: just over 0.1 m movement during this period (Fig. 12). The timing of this apparently static watertable corresponds to the construction of the drainage scheme and it is useful to closely examine immediate post-drain watertable responses.

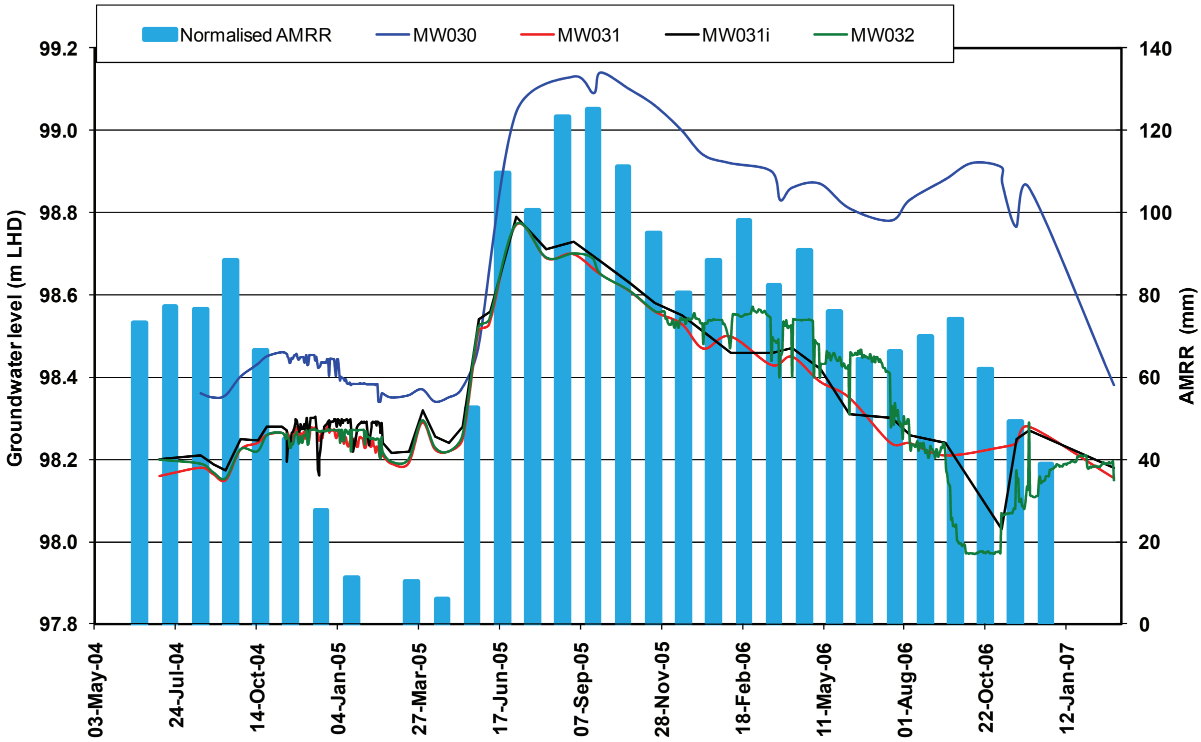

Figure 12 Comparison bore hydrographs

Transect 1

Bore transect 1 is perpendicular to the tributary drain and roughly parallel to and 350 m from the main drain (Fig. 11). It has 001 and 001d closest to the drain and extends to the footslopes and 006 at 400 m. Being fairly close to the main drain, the watertable could be partially affected by both the main and tributary drains.

There is an immediate watertable response to drain construction in the bores 20 m to 175 m from the drain, identified by the steepening in the downward slope of their hydrographs (Fig. 13). There are no similar responses in the two bores closer to the footslopes; their groundwater trends closely resembled those of the comparison bores.

There was no immediate watertable response as a result of the 950 m drain extension in mid-September 2005. This drain passes between 004 (175 m) and 005 (275 m), within 35 m of 005 at 240 m from the original drain. This completion of the drain extension resulted in 001–004 being completely enclosed by drains (Fig. 11) possibly resulting in the more noticeable watertable declines compared to the outer bores, after September 2005.

The overall watertable responses to the drains first appear insignificant, but comparisons with 031 show the following:

- From April to July 2005 the water level beneath 030 rose more than 0.60 m above its December 2005 (pre-drain) level (Fig. 12) while the maximum rise along transect 1 was about 0.3 m measured in 003 (100 m).

- The groundwater trend for 030 for the entire period of measurement (Fig. 12) is a 0.19 m rise while for the transect 1 bores in the same period all of the groundwater trends are downward 0.10–0.82 m.
- The watertable decline for the period of drain construction to the end of record was greatest in 001, 20 m from the drain and least in 006, 400 m from the drain. None of the comparison bore water levels showed a similar decline.
- The potentially drained watertable nearer the drain fluctuated more widely in correspondence with AMRR than either the outer (275 & 400 m) or comparison bores, reflecting filling and draining of the upper aquifer.

The limited pre-drain groundwater measurements show the water level of the deep aquifer at or below the watertable (Fig. 13). This indicates that at least during the later part of winter 2004 the site was a recharge area. The rapid response of the shallow watertable to the drain resulted in the reversal of this condition. Post-drainage, the site consistently became a discharge area as the watertable expressed in 001 fell and remained below the water level in 001d. Although the groundwater level did not closely follow that of the watertable (as at the comparison site), this presented a declining trend of 0.47 m.

The February 2005 water level fluctuations most evident at 275 m from the drain correspond to purging the bores and maintenance of the automated water level recorders. In all cases these activities resulted in immediate subsequent falls in watertable heights recorded in both manual and automated measurements. Although the causes of these watertable and bore responses require further investigation they do not affect the conclusions from this project.

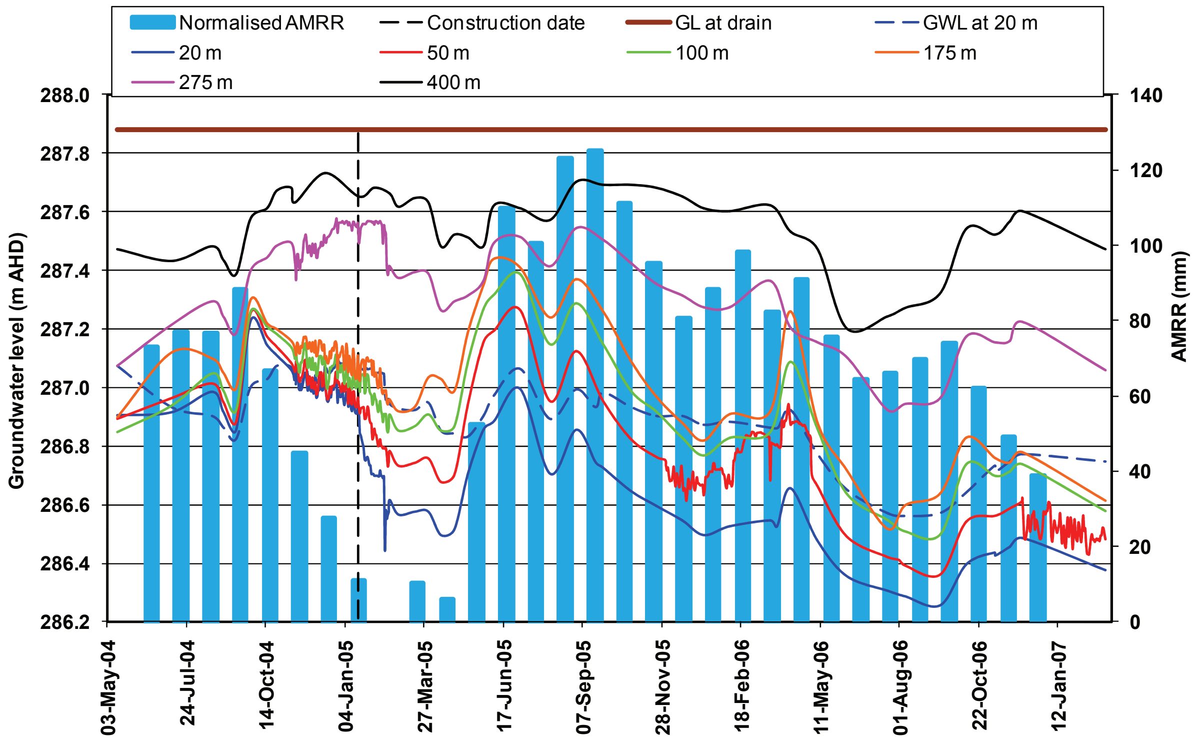

Figure 13 Transect 1 bore hydrographs

The 18 January 2005 pre-drained watertable closely followed the surface topography (Fig. 14). One month after drainage (19 February 2005) the decline in the watertable was greatest next to the drain, reducing with increasing distance. At 400 m there was no evidence of change.

Higher than monthly average May–June 2005 rainfall (Fig. 2) caused the watertable to rise to its maximum post-drain level in August 2005. The watertable rose nearly 0.5 m above the pre-drained level within 200 m of the drain, with limited response beyond. A further 400 days elapsed before the watertable fell to its minimum level on 7 August 2006 (Fig. 14). It was approximately 0.5 m lower than the pre-drain level and the average hydraulic gradient of the watertable towards the drain had increased by 16%.

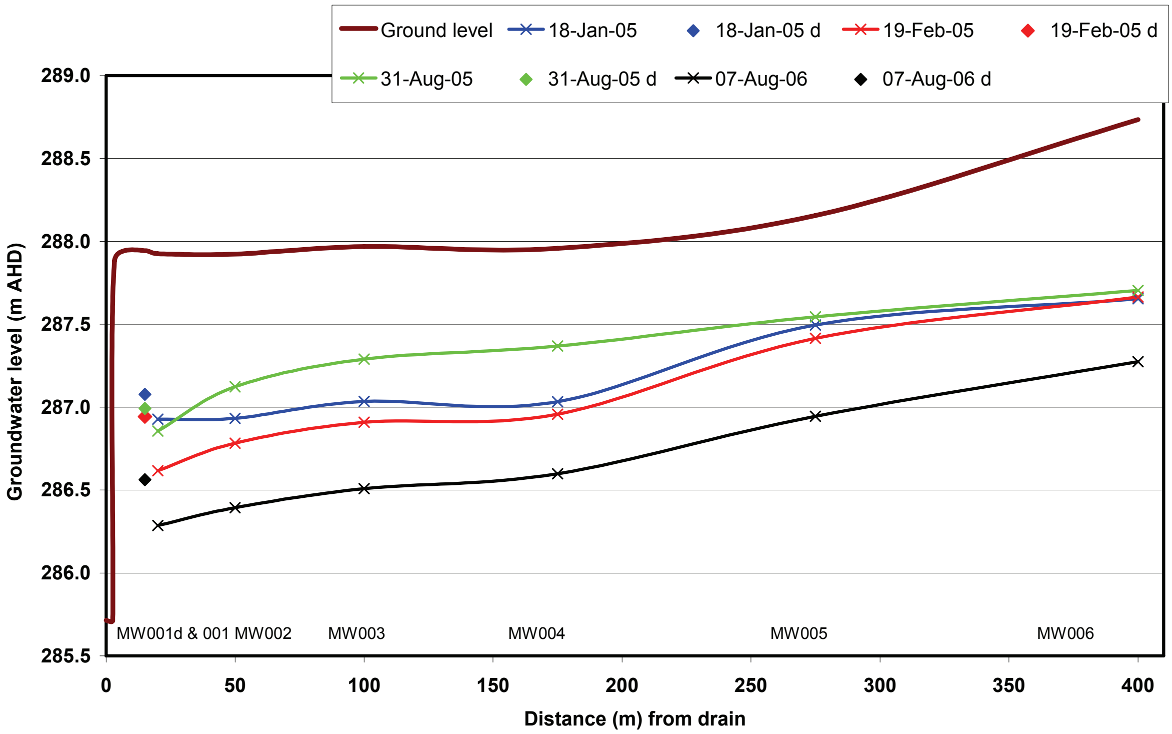

Figure 14 Transect 1 watertable profiles

Transect 2

Transect 2 extends SE from the main drain, downstream of the tributary drains and includes bores 007, 007d to 012. The land surface within 150 m of the drain is flat before sloping upwards at about 0.25%.

The immediate watertable response to the drain was almost identical to that of transect 1 and clearly detectable in 010 at 175 m (Fig. 15). Although there is evidence of some response to the drain beyond this distance, the clarity of the hydrographs is affected by possible interactions between response time lags and other variables such as barometric pressure-induced water level changes.

Water levels in this transect closely followed monthly AMRR trends showing the capacity for groundwater to move through this system quite rapidly in response to external and climatic influences. Most noticeable is that, although the post-drain January–June 2005 watertable reductions correspond to the AMRR trend, these responses are not reflected in the comparison bores.

All the hydrographs produced post-drain downward trends: the steepest represents a groundwater reduction of 0.43 m in 007, 20 m from the drain. The downward watertable trend was about 0.13 m for all the remaining bores. These uniform trends correspond more closely to the decreasing AMRR for the period of analysis (Fig. 4). This differed from transect 1 where the watertable decline was steeper nearer the drain than further away.

When compared to the comparison bores, the watertable up to 275 m from the drain appears to be at least 0.2 m lower than expected. By June 2006, the watertable measured within 275 m of the drain had fallen to 0.2–0.4 m below its pre-drain level (Fig. 15). The comparison bores showed June 2006 water levels still at or above those measured pre-drain (Fig. 12). The varied responses in the hydrograph for 012 at 400 m (Fig. 15) were caused by temporary damage and silting of the bore and so less frequent measurements from June 2005 to May 2006.

The water level in 007d declined 0.25 m during the monitoring period. In comparison bore 031i there was no overall decline although significant water level fluctuations were measured (Fig. 12). As in transect 1, constructing the drain appears to have caused the watertable to fall below the groundwater level but in this case the hydrograph of water level (007d) much more closely mirrored that of the watertable in 007. This was most evident at the time of drain construction when the water level fell 0.2 m within seven days, matching the decline of the watertable.

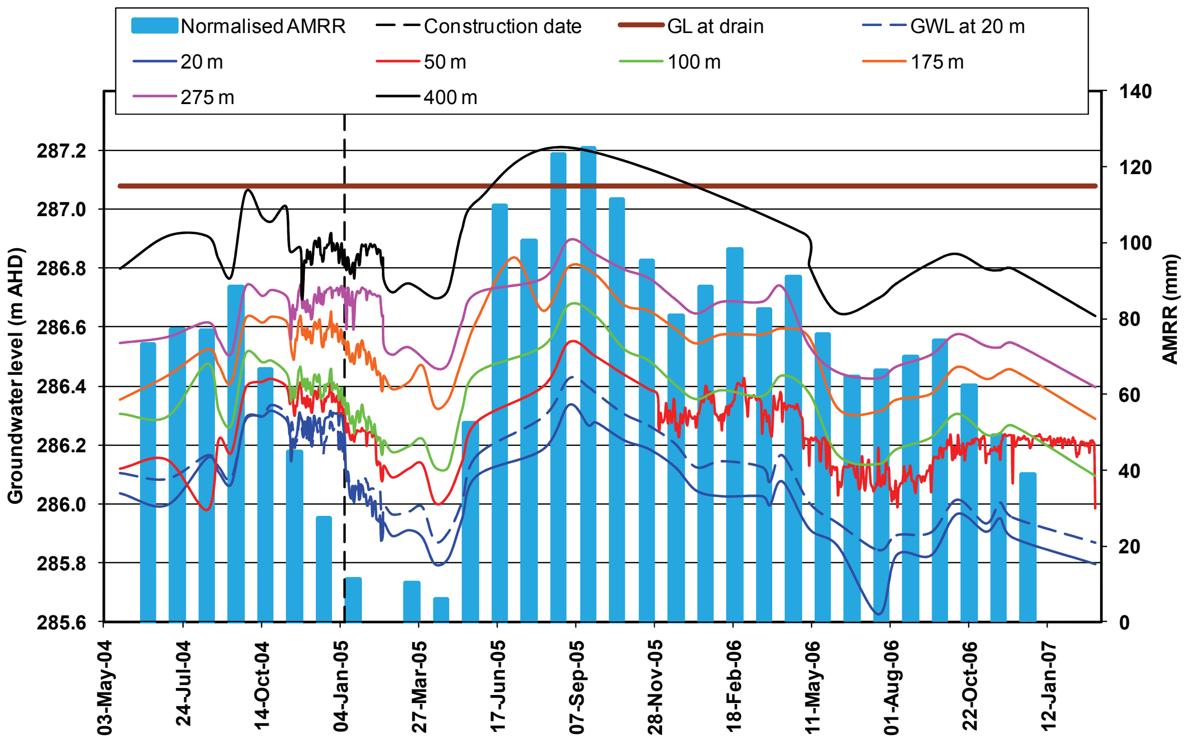

Figure 15 Transect 2 bore hydrographs

Watertable fluctuations were fairly uniform in height along transect 2 except within ~50 m of the drain (Fig. 16). The range of fluctuations was narrower than those of transect 1 and the watertable remained within 1.2–2 m above the drain floor at 20 m. Within one month of drain construction it is apparent that the watertable had fallen quite uniformly within 400 m by about 0.25 m.

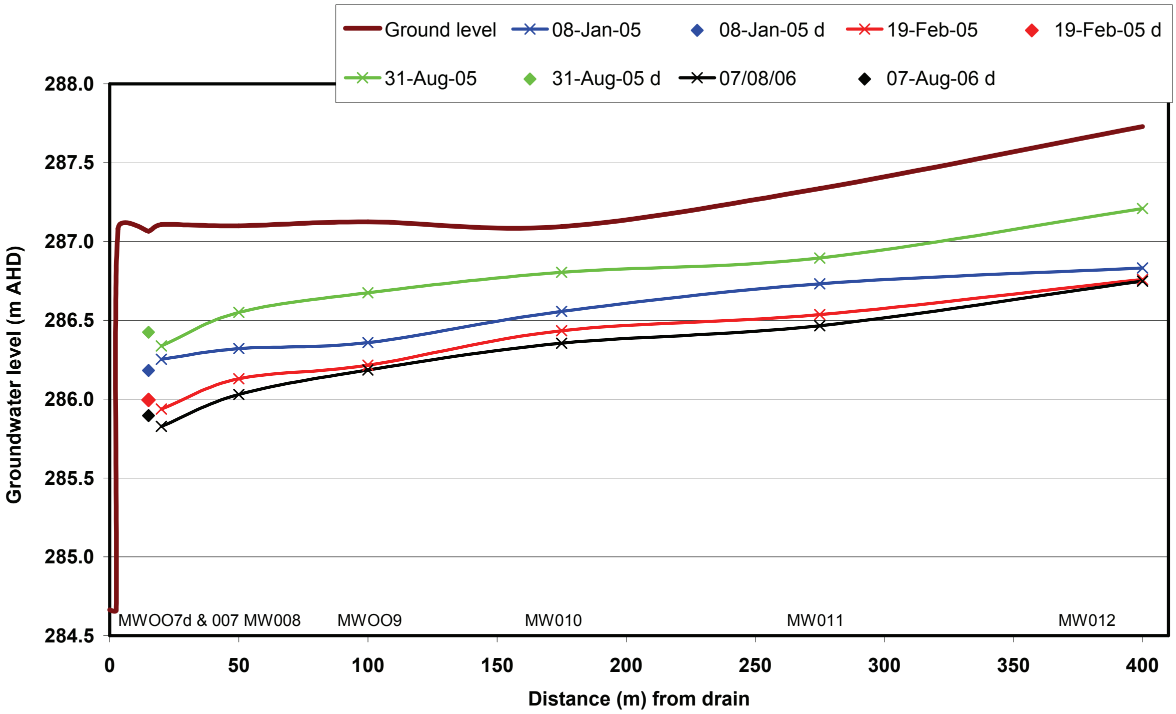

Figure 16 Transect 2 watertable profiles

Transect 3

Bore transect 3 is 800 m upstream from the discharge end of the drain, extending NW from its northern side (Fig. 11). Bores 013 and 013d are 20 m and 018 400 m from the drain and at the outer end of the transect. The topography along the transect rises from the drain at an average gradient at less than 0.1% (Fig. 17). At 100 m beyond 018, the slope of the valley flank is 0.5%, gradually rising to more than 1.0%.

The pre-drain watertable was close to flat (14 December 2004) with 0.1 m of fall from 018 towards 013 (Fig. 17). The watertable level in 013 fell 0.45 m within one month of digging the drain past transect 3. This fall reduced along the transect to 0.25 m at 400 m in 018. The greater watertable reductions in the bores closest to the drain resulted in an overall steepening in the slope of the watertable towards the drain by 18 January 2005. Throughout the subsequent range of fluctuations the average slope of the watertable towards the drain was consistently steeper than that of its pre-drain condition.

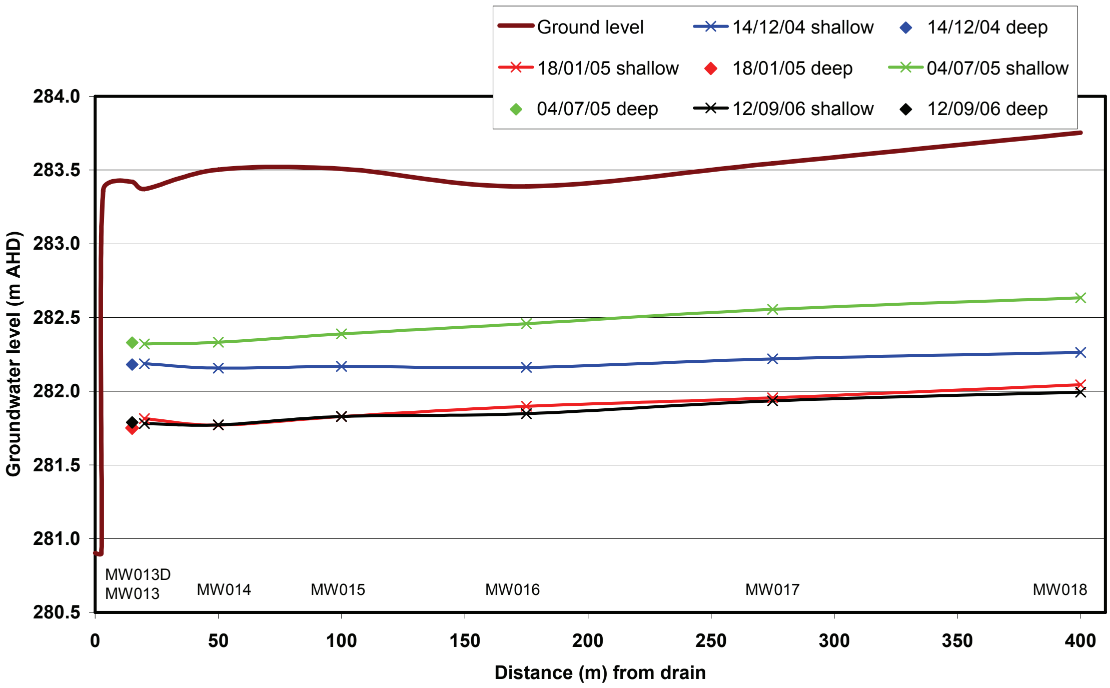

Figure 17 Transect 3 watertable profiles

Drain construction resulted in a sharp decline in the watertable noticeable at up to 275 m from the channel. Although this decline corresponded with a period of declining AMRR, it is believed to be primarily attributable to the drain (Fig. 18). There was no declining watertable trend associated with the downward AMRR trend between September–late December 2004, or seen in the comparison bores (Fig. 12). Within two weeks of drain construction the watertable had fallen by 0.45 m close to the drain and 0.25 m at the outer end of the transect, as discussed above. There was no significant difference between the water levels of the deep and shallow bores at 20 m from the drain.

For the remainder of the post-drain monitoring period the watertable trend closely corresponded to AMRR while fluctuating within a range 0.8–1.7 m above the drain base. Although it showed a falling trend from the July 2005 maximum level, this was punctuated by significant short-term fluctuations most noticeable in the bores within 100 m that corresponded with water-level changes in the drain (Fig. 18).

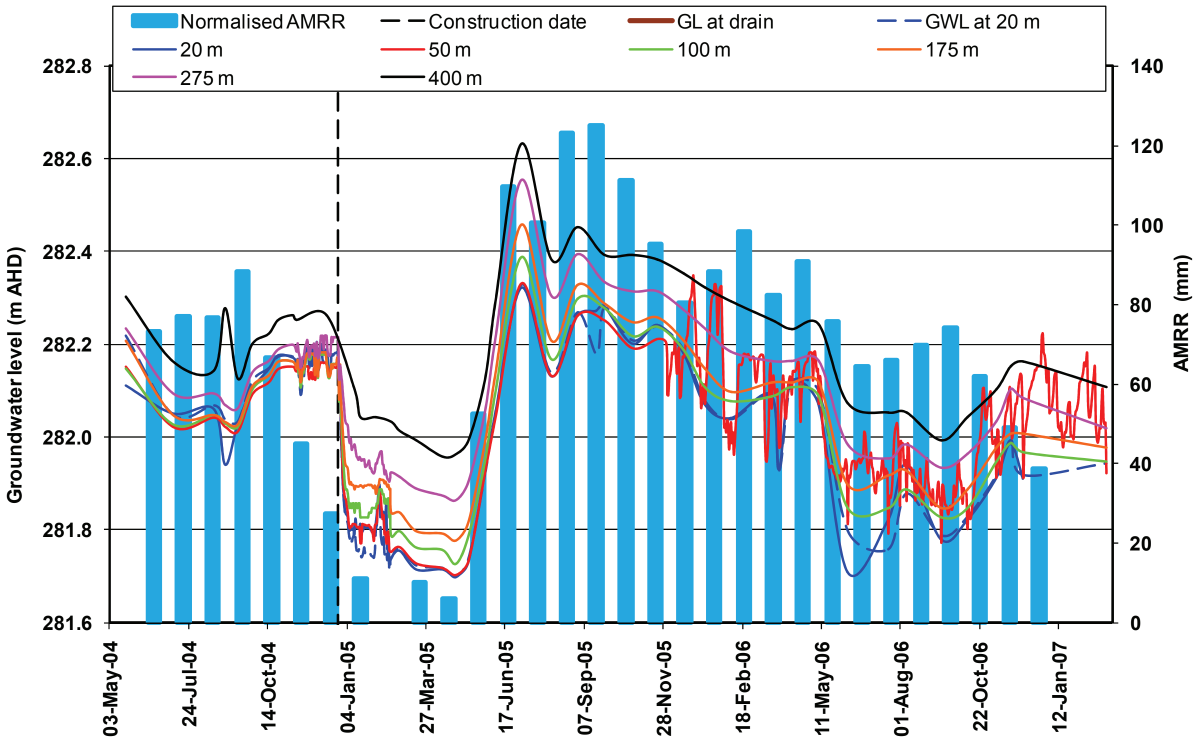

Figure 18 Transect 3 bore hydrographs

Stored water backed-up in the drain channel prevented the continuous inflow of groundwater needed to lower the watertable alongside this section of drain. When the sump and end of the drain were full to ground level, water backed up in the drain channel to nearly 400 m east of Holders Road (Fig. 11). The drain floor adjacent to Transect 3 is at about 280.9 m AHD elevation and 0.5 m above the drain floor entering the sump. When the water level reached ground level at the sump it was 2 m deep in the channel alongside transect 3, with a water surface elevation of nearly 282.9 m.

The rising and falling water level in the drain contributed to corresponding short- and long-term fluctuations in the watertable alongside (Fig. 19). The changes in elevation of the water level in the sump (606) and channel had almost immediate effects on the water level in bore 014 at 50 m. A rise in the water level in the channel to above that of the surrounding watertable caused a corresponding rise in the watertable. It is not known whether the watertable rises were caused by direct leakage from or the backing-up of groundwater inflow towards the channel.

A fall in the water level in the drain caused the watertable alongside to fall with 014 showing similar watertable level responses to those measured immediately after drain construction (Fig. 18). The responsiveness of the other bores in this transect to backwater is not known due to the lack of coinciding daily water level measurements for both bores and sump.

The lower water levels of both the drain and 014 reflect the greater emphasis on pumping from the drain May–October 2006. Equipping the pump with an automatic starter and water level sensor in the sump resulted in the water level being maintained an average ~0.4 m below its previous levels. This lower sump water level trend was reflected by the average water level in 014 falling by ~0.2 m. The return to manual pump operation in October 2006 saw both the water level trends of the sump and 014 return to their pre-May 2006 conditions (Fig. 19).

The water level trends in the sump and 014 are reflected in 018 at 400 m from the drain (Fig. 19). With an average May–October 2006 water level reduction of about 0.15 m, 018 appeared to have responded least to this period of enhanced pumping (Fig.18).

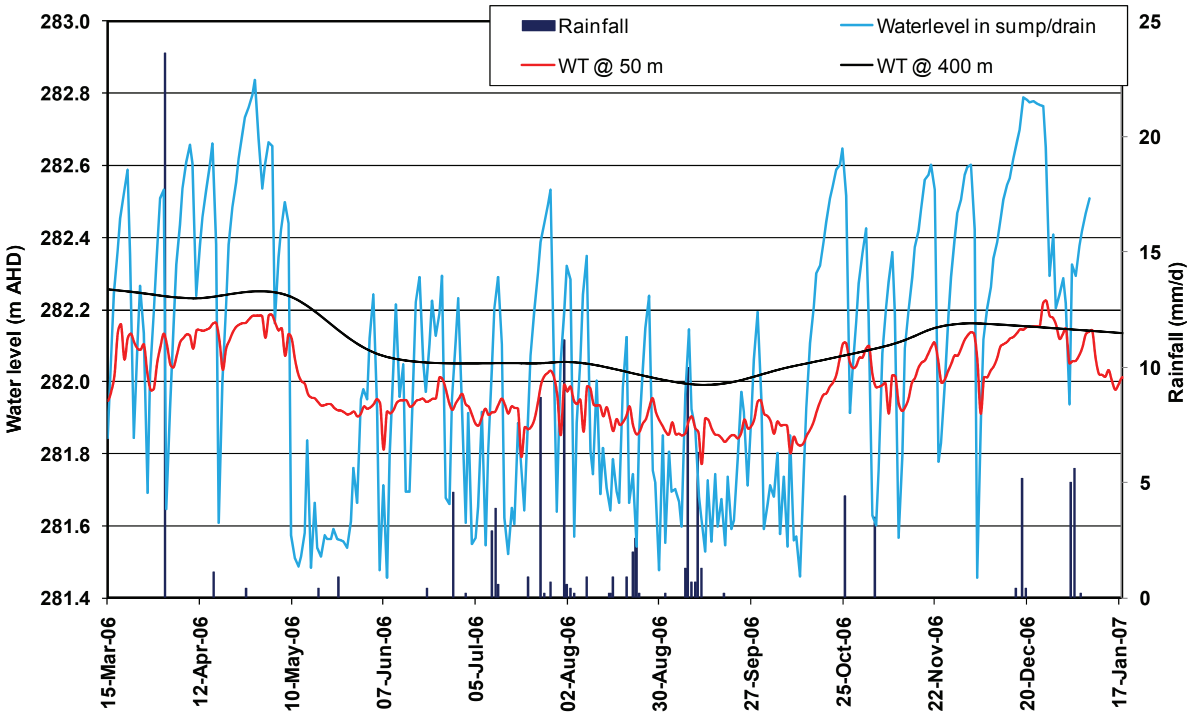

Figure 19 Transect 3 watertable partially responsive to transmission loss from the drain

Transect 4

Transect 4 extends NW from the middle of the basin cut-off drain and includes bores 019 and 019d at 20 m from the drain through to 024 at 400 m (Fig. 11). Two hundred metres beyond 024 the valley flanks start to rise to the NE. The land surface along the transect is flat, rising by 0.25 m along its length with a shallow depression that can become inundated (Fig. 20).

The pre-drain (13 December 2004) watertable sloped towards the cut-off drain with an average hydraulic gradient of 0.2 m along the 400 m (Fig. 20). By 27 December 2004, two weeks after drain construction, the watertable had fallen to its lowest level and its average gradient towards the drain increased to 0.45 m per 400 m. The watertable had fallen by 0.3 m in 019 at 20 m from the drain and 0.2 m in 023 at 275 m.

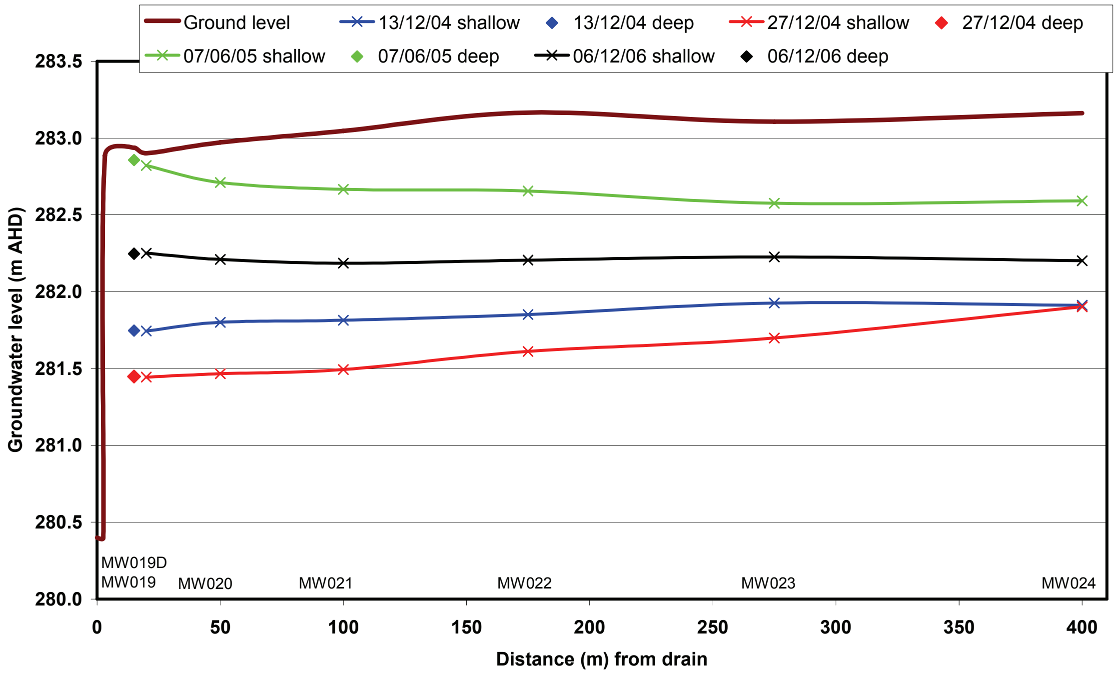

Figure 20 Transect 4 watertable profiles

Rapid watertable falls in response to drain construction were seen in all the bores but 024, before watertables start rising in response to leakage from the basin (Fig. 21). The lack of response in 024 was partially due to infrequently measuring this bore. In the period between measurements for 024 (27 December 2004 and 18 January 2005) the watertables in the other bores had fallen and risen again to their pre-drain levels.

The gradual filling of the basin caused leakage back into the cut-off drain, increasing its filling frequency and duration. As for transect 3, the water level rises in the sump and drain reflected rises of the adjacent watertables. On occasions, the water levels measured in the bores closest to the drain were higher than in bores further away (Fig. 20). Under these conditions the gradient of the watertable was away from the drain and basin indicating that groundwater was being recharged from basin, sump and drain leakage.

In June 2005 the cut-off drain was disconnected from the main drain (Section 3.2) and remained filled with basin leakage and groundwater. From June onwards, the water level in the cut-off drain reflected the groundwater level surrounding the basin, enabling leakage from the basin to flow uncontrolled through the cut-off drain and be reflected as a watertable rise in the surrounding landscape.

The cessation of pumping from the cut-off drain is not thought to be the only contributor to the 0.7 m watertable rise surrounding the basin in May–June 2005. During this period AMRR increased more than 100 mm, primarily due to the two > 20 mm May 2005 rainfall events (Fig. 2).

The maximum watertable height on 7 June 2005 (Fig. 21) coincided with the periods of maximum water level in the basin (Fig. 23), highest potential recharge and lowest potential evaporative loss. During these conditions a groundwater mound developed around the basin as represented by the watertable gradient reversal from towards to away from the cut-off drain, with the highest watertable closest to the drain (Fig. 20).

The watertable profile suggests that the groundwater mound was within 275–400 m of the cut-off drain which is similar to the extent of the noticeable drain responses in the other transects. Although the watertable rose nearly 0.7 m from its pre-drain level, this appears to be mostly associated with natural increase as measured in the comparison bores at the time. During the same period the groundwater level in 030 rose more than 0.7 m (Fig. 12) and the watertable in transect 3 more than 0.6 m (Fig. 18). The decays of these watertables also closely coincided with each other although the groundwater mound adjacent to the cut-off drain persisted on 6 December 2006 (Fig. 20).

As in transect 3, the watertable and underlying aquifer responses suggest an unconfined groundwater system to at least 20 m depth in 019d. Neither drainage responses nor watertable rise driven by leakage from the basin appeared to result in any difference in the water levels between the deep and shallow bores 019 and 019d (Fig. 21).

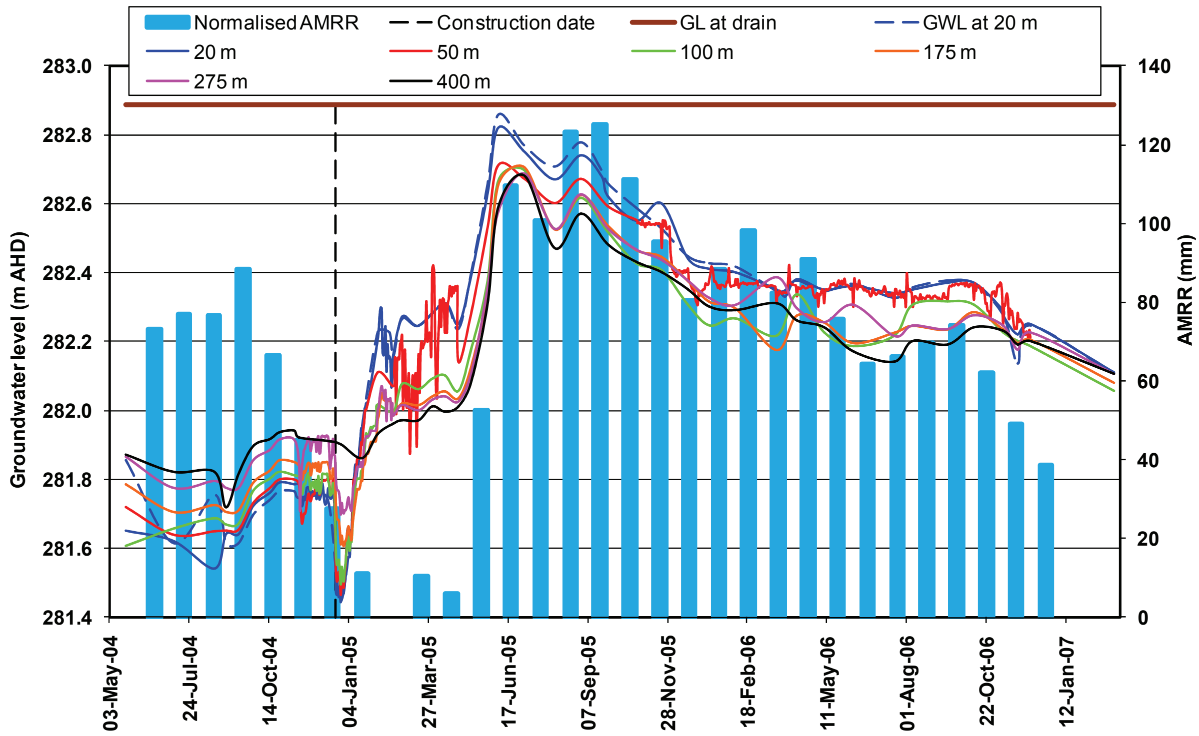

Figure 21 Transect 4 bore hydrographs

Transect 5

There are several bores around the outside of and downstream of the basin, including those of transect 4 (Fig. 11). Transect 5 consists of an alignment of selected bores that starts from 025 on the NE outside of the cut-off drain and runs through the basin WSW to 035 on the neighbouring property (Fig. 22). The 1080 m transect includes 026 (in the centre wall of the basin) and 027, 027i and 027d, 50 m to the WSW. The four downstream bores 033, 034, 35 and 039 were part of the group of 10 drilled later in the project (Section 3.3) and measured from 22 November 2005.

Groundwater and basin water levels are superimposed on the land surface profile along transect 5 which includes the cross-section of the basin and the cut-off drain at its NE. The pre-drainage and basin filling groundwater condition is indicated by the label ‘13/12/04’ (Fig. 22). At that time the watertable, intermediate and deeper groundwater levels (027, 027i and 027d) were all at the same elevation.

Not surprisingly, the level in 026 in the central wall of the basin rose to closely correspond

with the basin water level (Fig. 23). The 0.5 m groundwater level rises in 025 and 027 during January also corresponded closely with the basin filling. From then to late May 2005, recirculation of leakage by pumping from the cut-off drain appears to have prevented further watertable rise outside of the basin.

The water levels in the basin, cut-off drain and surrounding watertables peaked around 4 July 2005 (Fig. 23). These water levels were measured shortly after pumping from the cut-off drain stopped. In combination with the onset of winter rainfall, this caused an additional 0.25 m groundwater rise in 025 and 027 (Fig. 23), bringing their water levels to within 0.5 m of the ground surface at about 283 m elevation (Fig. 22).

The hydrograph of 028 (not shown) showed a steadily rising level from filling the basin to 4 July 2005. The level rise from 1.37 m below to 0.12 m above ground level corresponded with groundwater visible as pools on the surface adjacent to the basin wall (Fig. 48). Bore 028 is 50 m from the centre SSE side of the basin where there is no cut-off drain (Fig. 11).

The basin water level fell quickly after 4 July 2005 when the stored water was released into the larger contingency basin. For the rest of the project the basin water filled mainly the borrow areas and rarely rose above the natural ground level (about 283 m AHD; Fig. 10).

When compared with the hydrograph of 030 (Fig. 12), the effect of basin leakage on the surrounding watertable did not appear to be significant and was limited to some earlier and slightly increased localised watertable rises compared with what would have occurred naturally. Comparing the hydrographs of 025 and 027 (Fig. 23) with 030 (Fig. 12), some commonalities include:

- By 31 August 2005, all of the hydrographs had peaked at about 0.75 m above the pre-drain period height.
- By 6 February 2006 all of the hydrographs had decayed by nearly 0.25 m.
- After 8 November 2006 the rate of hydrograph decay increased significantly above its preceding trend.

On 22 November 2005 the watertable elevation surrounding the basin appeared more representative of that expected under natural conditions (Fig. 22) given its proximity to the centre of the valley floor. The watertable along the downstream 600 m from 027 more closely followed the surface topography than showed the development of a groundwater mound surrounding the basin. From November 2005 there was limited commonality between the hydrograph responses of the downstream bores and those closer to the basin (Fig. 23). The hydrographs of the downstream bores were more responsive to the AMRR while those closer to the basin were less responsive.

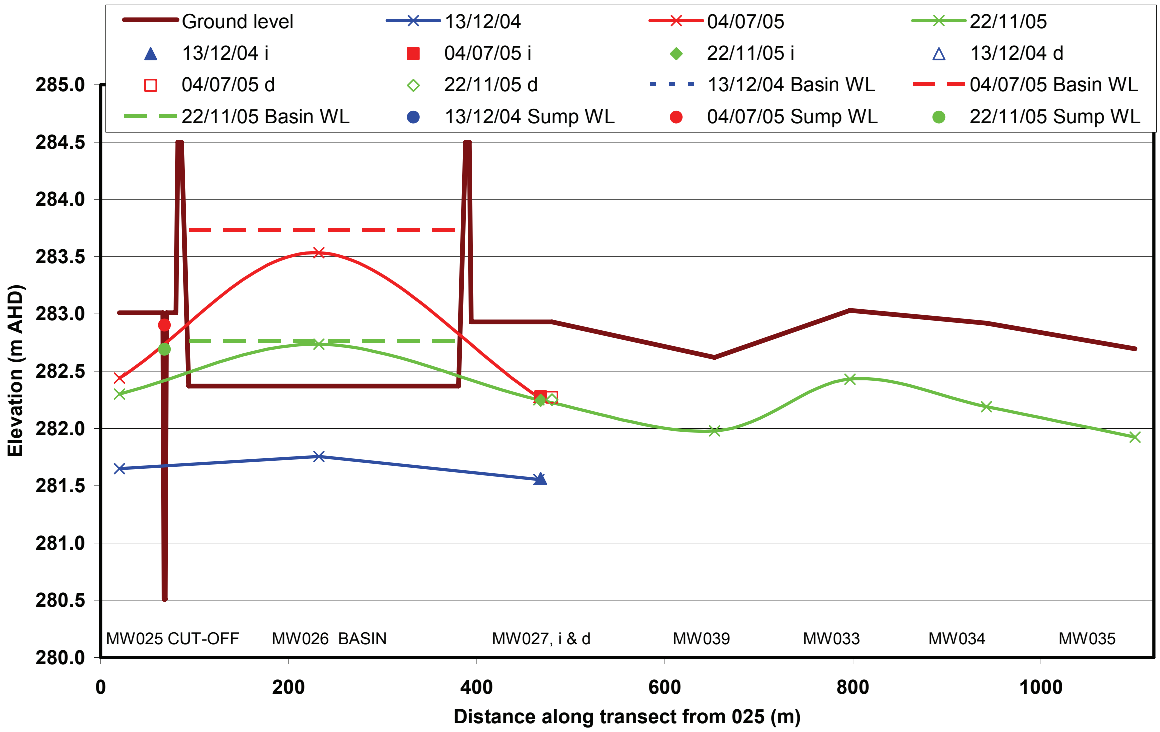

Figure 22 Watertable profiles for transect 5

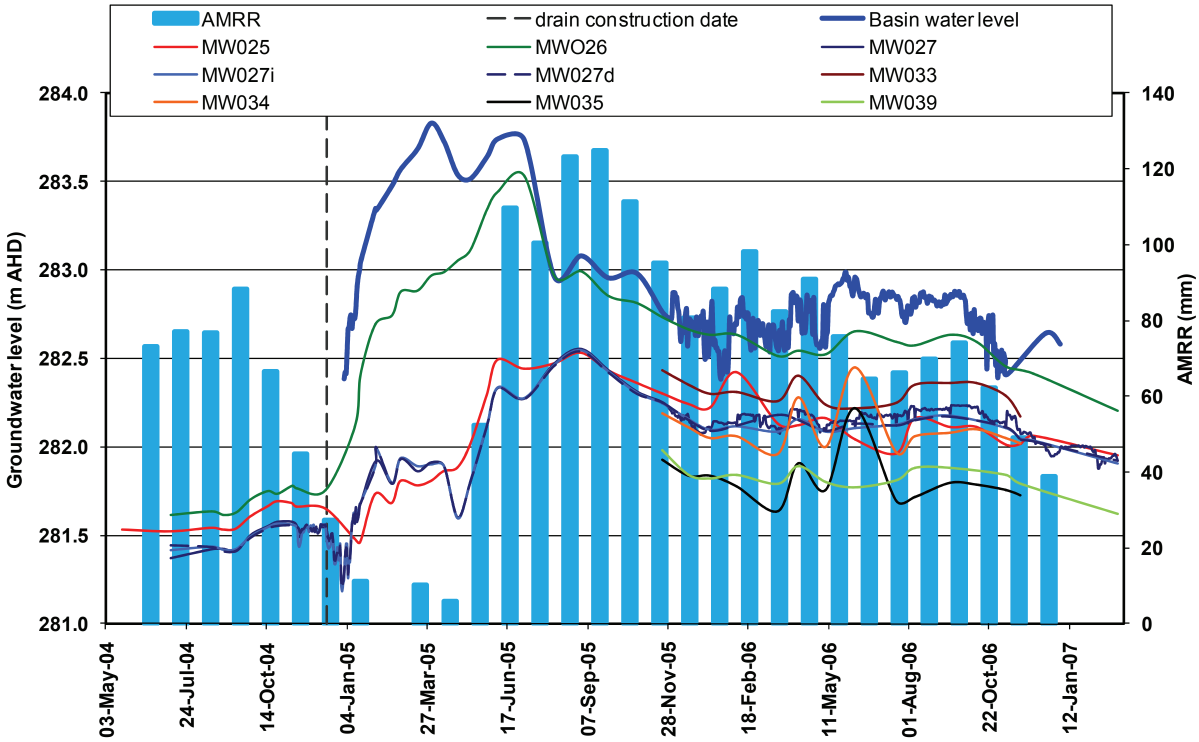

Figure 23 Transect 5 hydrographs

4.2 Groundwater salinity

Groundwater salinities generally increased with:

- closer proximity to the valley floor centre/drainage line
- depth
- distance downstream (east to west).

Groundwater salinities ranged from 5000 to more than 65 000 mg/L in the monitoring bores across the valley floor. A reducing watertable salinity trend from the centre of the valley towards the valley flanks (Section 2.6) is expressed by the average annual salinities of the bores in transects 1–4 (Fig. 24).

Deeper groundwater was on average about 5000 mg/L more saline than in the overlying watertable in each transect. There was a close relationship between the salinities and trends of the paired deep and shallow bores within each transect (Fig. 24), at the basin (027) and comparison site (031). Due to the lack of deep bores it is not known if deeper groundwater salinities decreased from the centre of the valley towards the valley flanks like the watertable salinities.

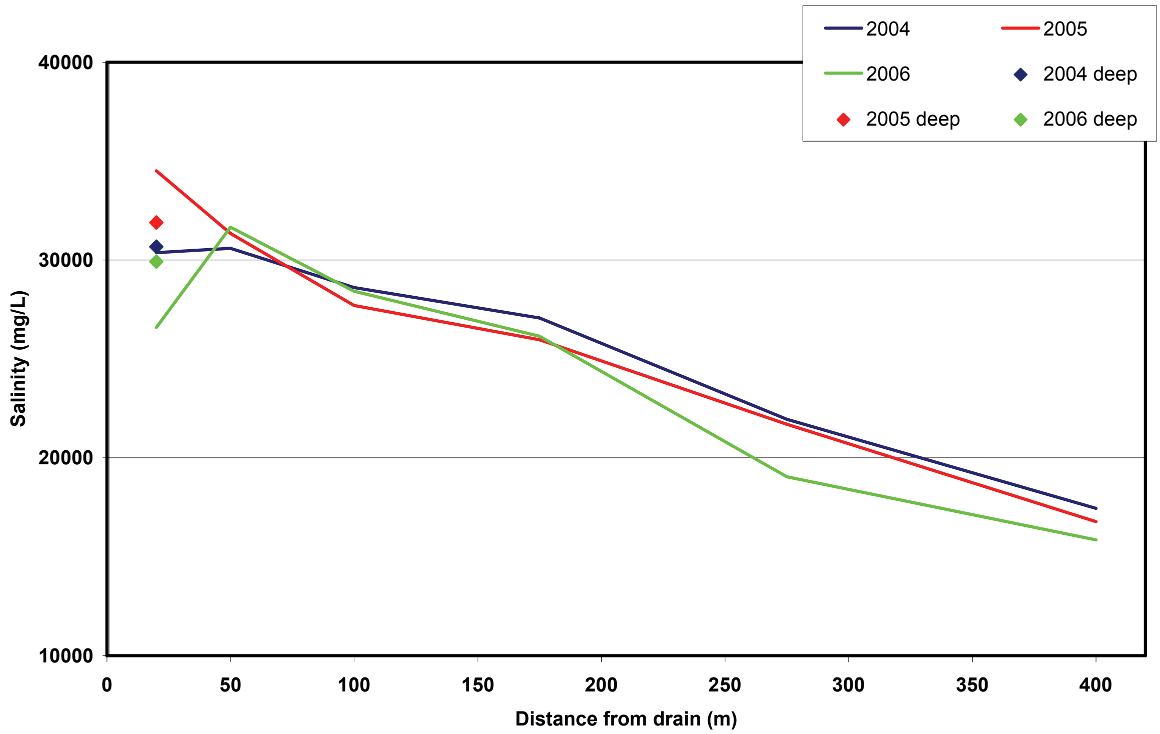

Figure 24 Average watertable salinity profiles and groundwater salinities for transects 1 to 4

Both watertable and groundwater salinities trended upwards from east to west, downgradient along the valley floor. These rising trends are best illustrated by the groundwater salinities measured in the transects 1–5 deep bores at or close to the centre of the valley floor (Fig. 25).

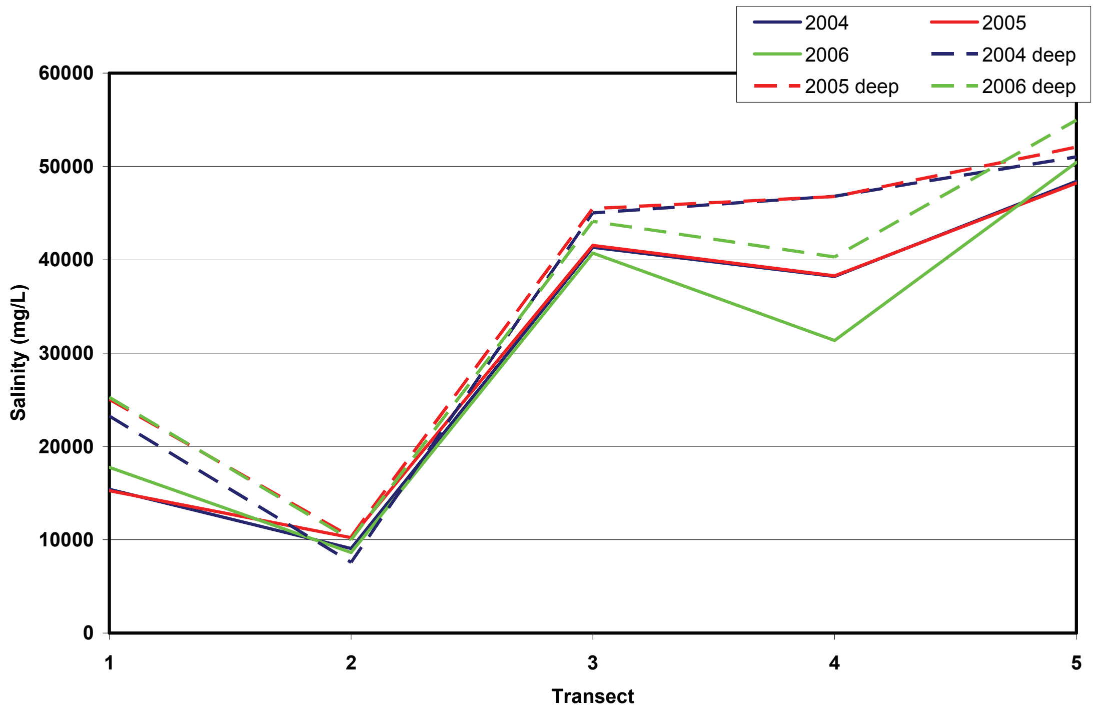

Figure 25 Average watertable and deep groundwater salinities for each transect

Some watertable salinities varied widely through the monitoring period, with the biggest change measured in bore 007 where, between April and November 2005, salinity increased from 6300 to more than 47 000 mg/L. These salinities were confirmed by a laboratory analysis and several in-situ measurements before they returned to near pre-increase levels. Adjacent bores displayed similar but smaller salinity fluctuations (Appendix CD 4.2b).

Watertable salinities partly reflected the extent to which each transect was encompassed by the poorly drained and salt-affected valley floor, and the underlying groundwater heads. Transect 1 extends upslope from the valley floor (Fig. 14) so the bores were likely to be less affected by runoff recharge and associated salt mobilisation. The land surrounding transect 1 was identified as a recharge area before drainage (Section 4.1) providing the conditions for some leaching of salts from the watertable to deeper groundwater.

To help distinguish seasonal trends from localised drainage effects the average salinities of shallow transect bores are shown in Figure 26. Together the charts suggest:

- A seasonal trend of increasing salinity around July–October each year could be caused by a rise in the deeper groundwater level. This is supported by the September peak (mainly in transect 2) coinciding with the highest watertable levels during the same period (Fig. 15). This upward trend was diminished in 2006, coinciding with lower rainfall and a smaller watertable rise.
- Watertable salinities were strongly influenced by seasonal variations, localised recharge, evapoconcentration and deeper groundwater salinity and level fluctuations. This is clearer in transects 1 and 2 where there was some contrast between the watertable and deep groundwater salinities. In these cases the fluctuations in the shallow and deep groundwater levels caused bigger fluctuations in watertable salinity than in bores further downstream.
- Watertable salinity trends were influenced by landscape position. The trends of transects 1 and 2 both close to the valley flanks are comparable as are the trends in transects 3 and 4 on the broad valley floor, although the response trends in these two transect groups are quite different.
- No watertable salinity responses could be attributed to installing and operating the drain or evaporation basin. There was no detected change in salinity beneath the basin probably because the salinities of the leaked water and of the underlying groundwater were similar.

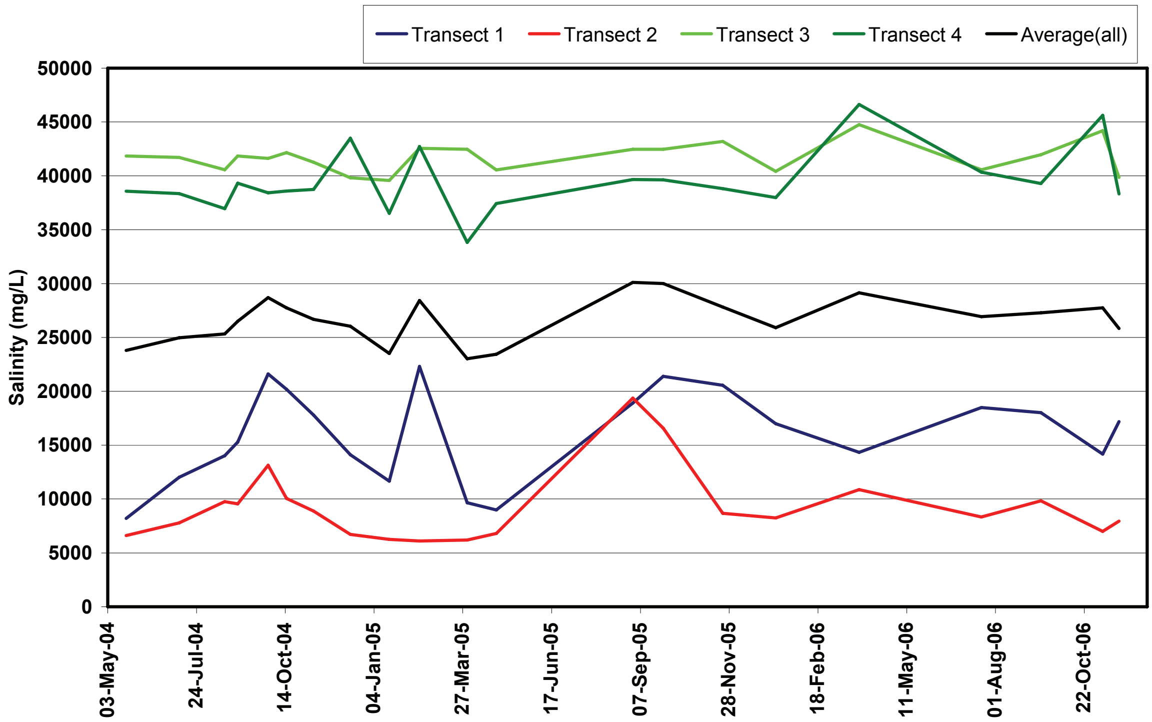

Figure 26 Average watertable salinities for transects 1–4 singly and combined (all)

The deeper groundwater salinities rarely fluctuated outside a 5000 mg/L range throughout the measurement period (Appendix CD 4.2b). Deeper groundwater salinities beneath the transects showed no deviation from long-term or seasonal trends that indicated possible responses to the drain (Fig. 27). Of note is that, except for transect 2, average watertable salinities (Fig. 26) only rose close to but never above those of their respective underlying groundwaters.

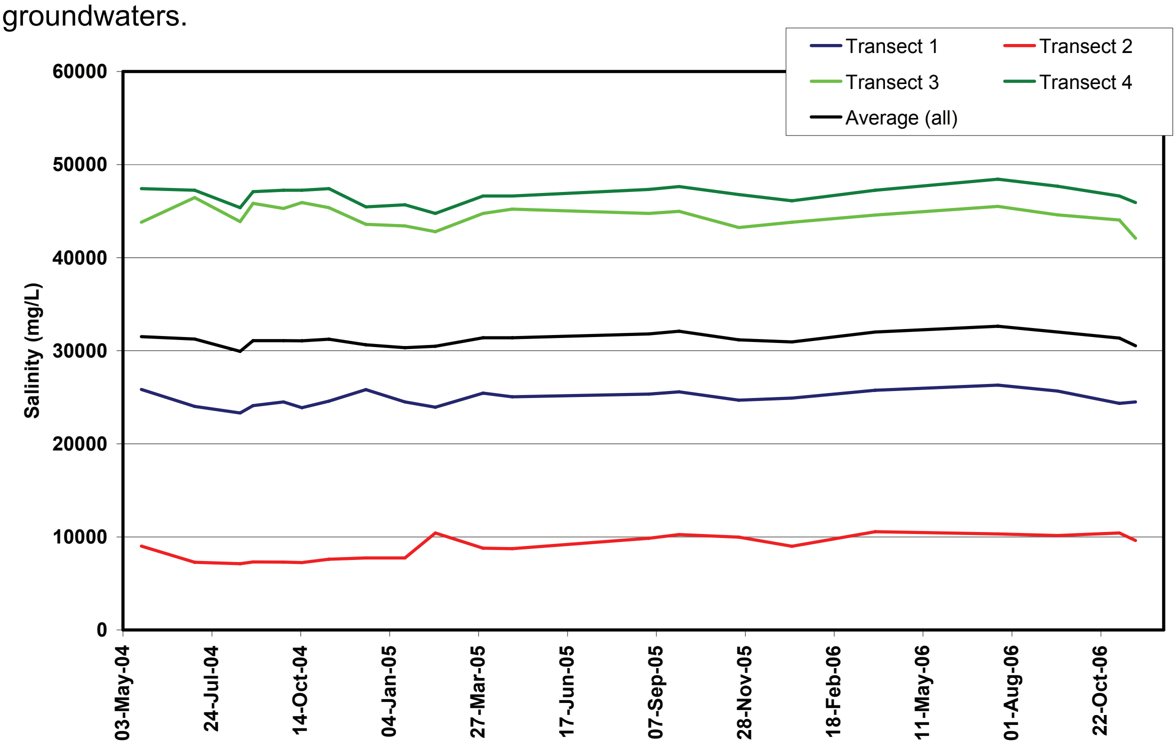

Figure 27 Average deep groundwater salinities for transects 1–4 and average of all results

The similar salinities of the drain, basin and groundwater made it impossible to detect groundwater salinity changes in response to drain or basin leakage (Section 4.1). The groundwater surrounding and downstream of the basin was, for the most part, more saline than the drain discharge but less saline than water held in the basin. Transect 5 watertable salinities remained and fluctuated mostly within the range 40 000–60 000 mg/L, with deeper groundwater salinities remaining close to 50 000 mg/L (Appendix CD 4.2b).

4.3 Groundwater pH

No consistent commonalities were evident between the pH trends of the groundwater measured in any of the bores. The average pH of all watertable measurements in transects 1–4 was 7.07 with a range 6.5–7.8 (Fig. 28). Individual deep and shallow groundwater values ranged from 5.5 to 8.6 (Appendix CD 4.0).

An apparent relationship between watertable height and pH may be spurious and related to the frequency of the measurements and the regular evacuation of the bores although average watertable pH levels falling towards the end of the winter (September–October) rainfall months (Fig. 28) coincided with the lowest water levels measured in many of the bores.

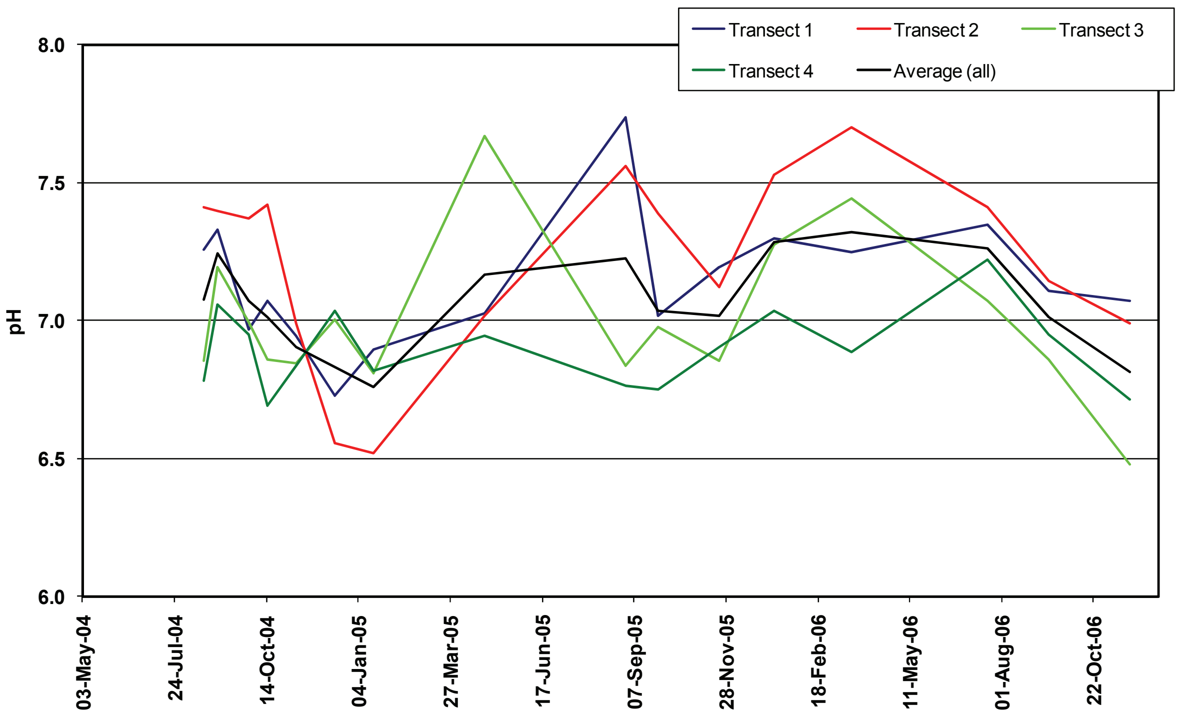

Figure 28 Average watertable pH of transects 1–4

Deeper groundwater pH levels were similar to those of their overlying watertables in transects 1–4, ranging from 5.9 to 8.4 (Fig. 29). As mentioned above, there was no consistent relationship between the pH trends of the deep and shallow groundwaters of even paired bores. Like the watertable pH, deeper groundwater pH reflected a spurious seasonal trend coinciding with rainfall and/or groundwater level changes.

Figure 29 Deeper groundwater pH for transects 1–4

Like salinity, watertable pH showed a falling trend from the centre towards the edges of the valley floor (Fig. 30). This trend reversed within 100 m of the drain alignment with the most significant fluctuations and decline in average pH values measured in 2006. At 100 m from the drain 2006 average pH levels were about 0.2 units higher than in 2004–2005 while those 20 m from the drain were 0.2 units below. The pH levels measured along transect 4 had the greatest influence on these anomalous 2006 values (Appendix CD 4.2b).

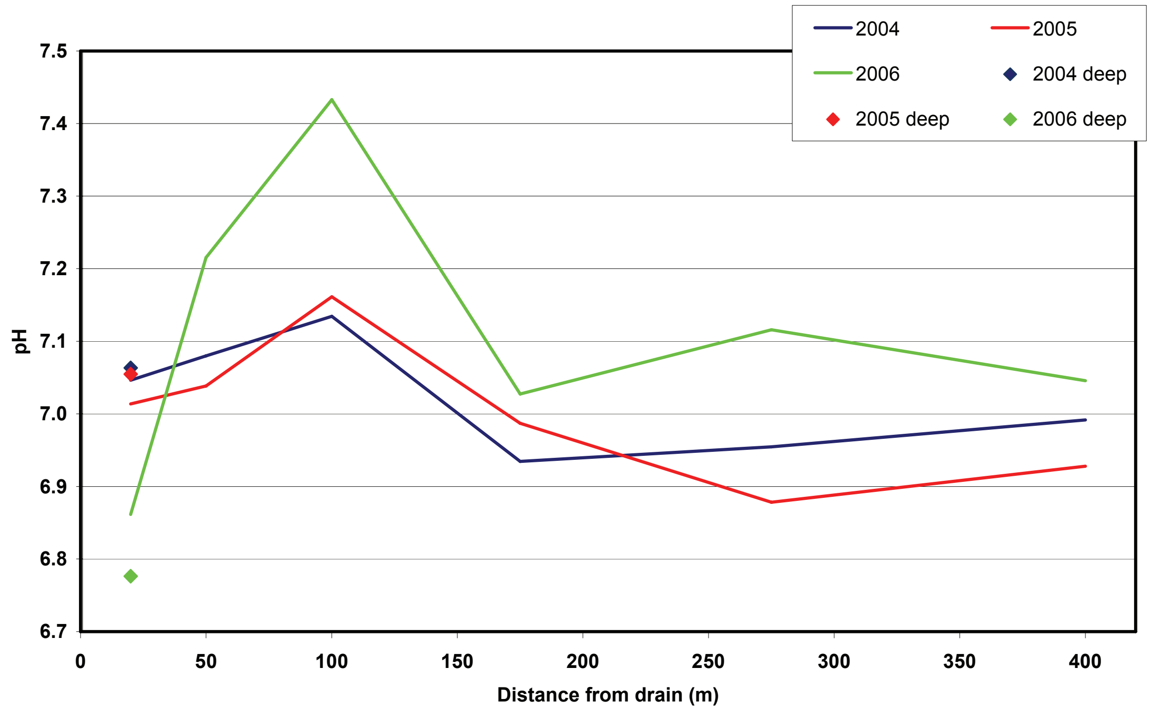

Figure 30 Average annual groundwater pH values along transects 1–4

Average groundwater pH values for 2004–05 showed no trend to a declining trend from east to west, downgradient along the valley floor (Fig. 31). During 2006 the declining pH trend increased mainly in response to increases in the average pH values for transects 1–2. Contrary to this, the 2006 pH trend of the deeper groundwater became one of increasing downgradient along the valley floor. The increasing trend coincided with falling deep groundwater pH values for transects 1–3, and increasing watertable pH values for transects 4–5.

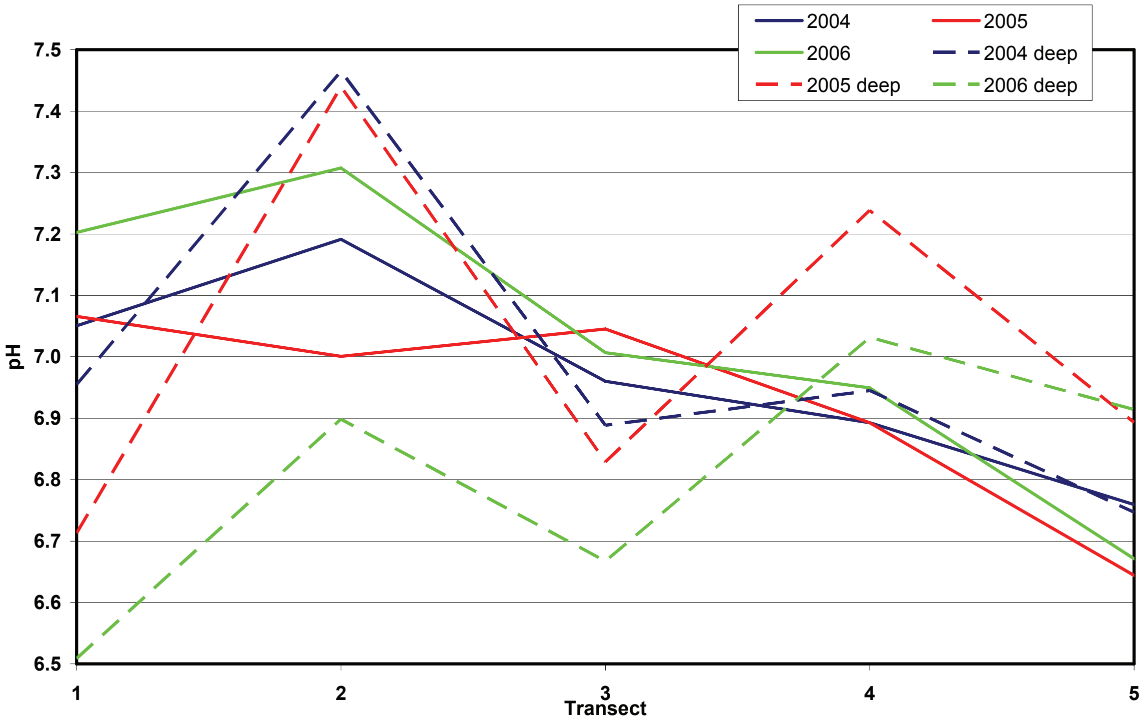

Figure 31 Average annual shallow and deep groundwater pH values of transects 1–5

4.4 Groundwater chemistry

The groundwaters were neutral (laboratory measured pH). Morawa groundwaters had extremely low concentrations of dissolved metals like iron and aluminium, trace metals and rare earths; in fact, they were often below detection limits (Appendix CD 4.4). These levels are thought be low because the groundwater pH was mostly above 6 and most of the trace metals and rare earths are insoluble at pH > 6. They are unlike groundwaters of the central and southern Wheatbelt where many drains produce acidic water (Silberstein et al. 2005) that may also have high concentrations of dissolved metals like iron, aluminium, trace metals and rare earths.

The groundwaters have a chemical composition similar to seawater with the major ion concentrations dominated by sodium and chloride, like most groundwater in the Wheatbelt. This suggests that the most important process controlling groundwater chemistry was the concentration of salts derived from marine aerosols, transported to the groundwater by rainfall recharge and concentrated by evapotranspiration, or even direct evaporation by capillary discharge from the watertable (Dogramaci & Yesertener 2001).

The total phosphate and total nitrate concentrations were very low in relation to the maximum levels acceptable in the Australian environment (Environmental Protection and Heritage Council 2000) for livestock drinking and ecosystem protection. Watertable phosphate (P) levels in any one bore varied from below detection levels to 1.2 mg/L, with most around 0.3 mg/L. Total nitrogen (N) levels ranged from 0.3 to 17 mg/L in both the shallow and deeper groundwater and higher in shallow groundwater and at the eastern end of the property. Cropping practices were probably the main source of N in the shallow groundwater and this distribution may reflect the retreat of agriculture from west to east ahead of the salinising landscape.

4.5 Main drain discharge

At the start of drain excavation a manually operated tractor power take-off centrifugal pump with a maximum capacity of 20 L/s (1730 kL/d) was used to lift the water from the sump to the basin. The pump's capacity was quickly exceeded as drain-flow increased when the drain was extended and the leakage recirculation rate from the basin increased. Just over a month after drainage started the pump was replaced with a 60 L/s (5200 kL/d) capacity one.

The pump discharge hydrograph shows the lower pumping rate from December 2004 to February 2005 (Fig. 32). By February 2005 the basin held approximately 13 000 kL in the western cell and the measured discharge rate from the main drain was highly modified by leakage recirculated from the basin and the frequency and duration of pumping from the sump into the basin. From 21 December 2004 to 15 June 2005 the water pumped from the sump into the basin consisted of drain-flow and recirculated leakage from the basin into the cut-off drain and sump.

On 10 February 2005 the installation of the larger pump (60 L/s, 5200 kL/d) is reflected by the increased daily pumping rate in the hydrograph, especially from February to June 2005 (Fig. 32). During this period the leakage recirculation rate from the basin increased, associated with its filling and the increase in head between the basin water and the cut-off drain. The average pumping rate from project commencement to June 2005 was 1293 kL/d (Table 1) and equivalent to 1.87 L/s for each kilometre of drain (including the cut-off drain).

The high leakage recirculation rate combined with two significant rainfall events in May 2005 (Fig. 2) led to expanding the basin and disconnecting the cut-off drain in June 2005: seen as an apparent decline of about 1000 kL/d per pumping event (Fig. 32) and 10 000 mg/L reduction in flow-weighted salinity (Fig. 34). At the time of disconnection in mid-June 2005 the average pumping rate was 1293 kL/d. For the remainder of 2005 the average pumping rate was 817 kL/d (Table 1) with the difference between the two rates reflecting an average 476 kL/d of possible leakage recirculated from the basin.

Disconnecting the cut-off drain from the main drain also led to an average 5500 mg/L (11.5%) reduction in the pumped flow-weighted salinity of the discharge (Fig. 32). The initial higher salinity reflected leakage of the more saline evapoconcentrated water from the basin into the cut-off drain. From December 2004 to June 2005 basin salinities ranged from 46 000 to 53 000 mg/L.

Over the 749 days of measurement, 553 000 kL of water at an average salinity of 42 500 mg/L were lifted from the drain into the basin. This equates to 1.07 L/s/km pro-rata drain length operational at different times during the project (Table 1). The total salt load of 23 600 t was lifted into the basin with the drain discharge.

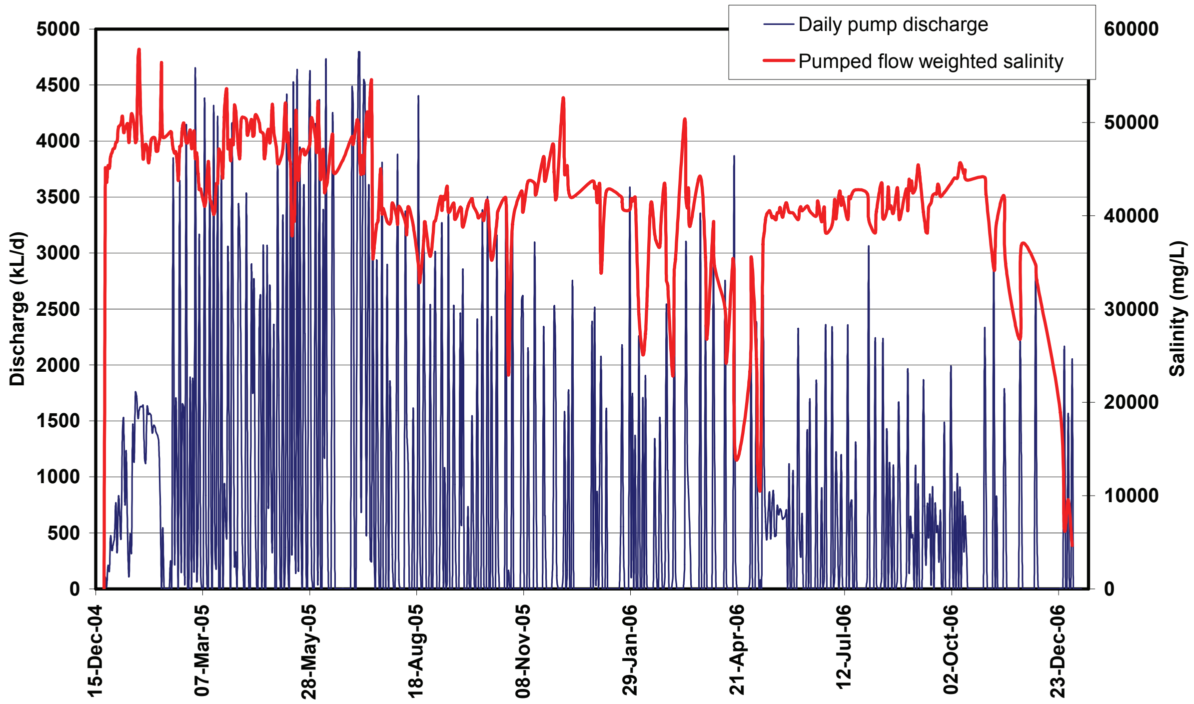

Figure 32 Pumped volume and flow-weighted salinity from the sump to the evaporation basin

Table 1 Main drain discharge summary

Dates	Volume			Average salinity (mg/L)
	Total (kL)	Average (kL/d)	L/s/km	
21/12/2004–15/6/2005	227 574	1293	1.87	47 460
16/6/2005–31/12/2005	161 848	817	1.22	42 010
1/1/2005–31/12/2005	366 167	1061	1.56	45 450
1/1/2006–31/12/2006	161 672	443	0.62	38 420
21/12/2004–10/1/2008	553 147	739	1.07	42 540

Drain-flow and basin leakage built up in the sump, cut-off drain and lower end of the main drain between pumping. The combined below-ground-level storage capacity of these structures was 10 500 kL. The sump was rarely pumped dry, ruling out the assumption that the volume of water pumped represented that accumulated since the previous pumping event. This means that the following estimation of actual drain discharge compared to pump discharge was compromised by:

- the recirculation of basin leakage at varying rates until 15 June 2005
- the carry-over of unknown volumes of water between pumping events
- the loss and possible return of water to and from the drain as the impounded water level in the sump, cut-off drain and lower drain rose and fell in relation to the surrounding watertable (Section 4.1). This continued for the duration of the project.

A forward-moving average of 1920, 15-minute pump-discharge volumes was used to represent drain discharge in place of pump discharge. This number of 15-minute intervals is equivalent to a 20-day forward-moving average. These intervals were used rather than the daily pump-discharge values to enable re-calculation of salinities and salt loads to correspond with the calculated discharge results.

The method did not succeed in screening out significant human-induced influences on discharge such as the pumping stoppages in February and June 2005 (Fig. 33). However, it revealed significant discharge trends in late 2005–06. From mid to end 2005, daily discharge reduced steadily from around 1000 to 500 kL/d. The 500 kL/d discharge was sustained through early 2006, possibly in response to the out-of-season rainfall in January and February (Fig. 2). There was only a small decrease in discharge in response to the below-average 2006 winter rainfall of May–October. By October 2006 discharge was decreasing noticeably, corresponding with decreasing AMRR and falling watertables (Section 4.1).

The human-induced impacts did not mask the close correlation between drain discharge and AMRR trend. Significant upward discharge trends were clearly evident in response to increases in rainfall in May 2005, August 2005, January 2006 and April 2006 (Fig. 33). The volumes associated with these increased discharges far exceeded those which could be generated by rainfall directly into the drain. Conversely, downward-trending discharge corresponded closely with declining rainfall with short-term falling trends most evident from August–December 2005 and from September 2006 (Fig. 33).

Both drain discharge and rainfall presented general declining trends from June 2005 which followed the period of high drain discharge December 2004–May 2005. This initial period of high discharge was associated with high basin leakage recirculation, the onset of winter rainfall (Fig. 2) and the apparent release of accumulated groundwater surrounding the newly dug drains (Section 4.1).

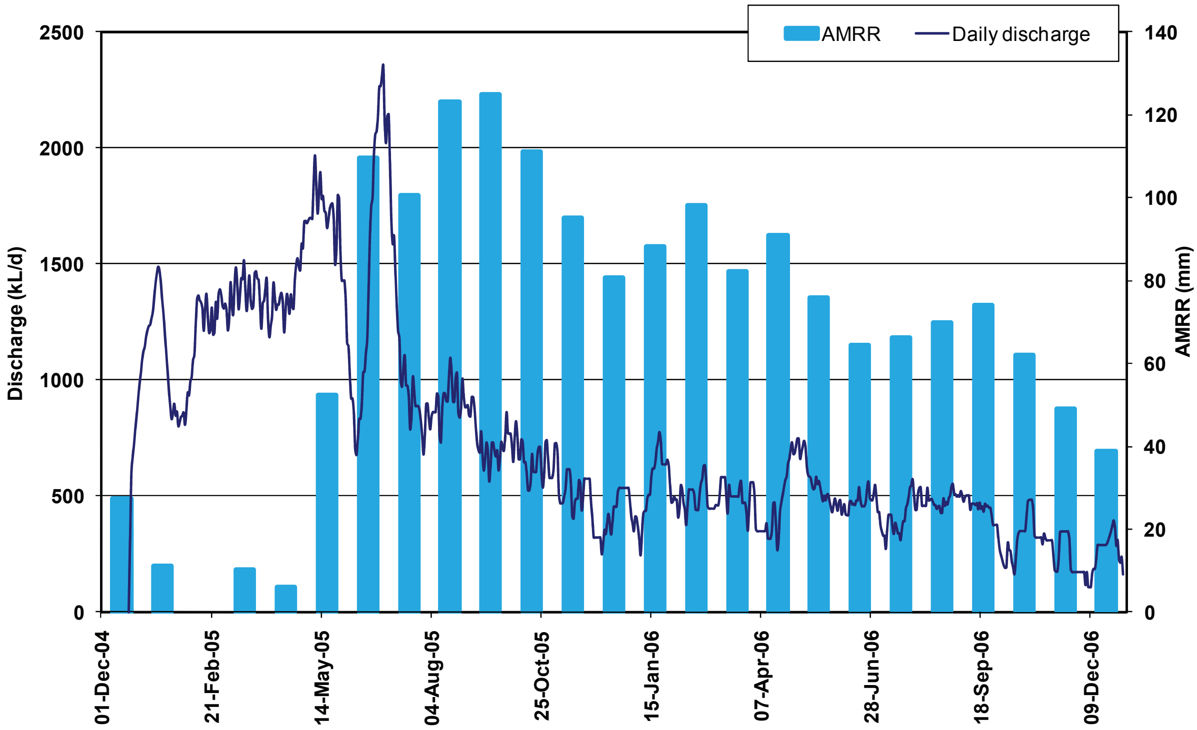

Figure 33 Main drain discharge and AMRR trend

Discharge reduced about 5000 mg/L in flow-weighted salinity from 16 June to 15 July 2005 after the basin leakage was isolated from drain-flow (Fig. 34). Other short-term salinity data trends are thought to be spurious, possibly related to the pump operation.

Throughout the project, the volume of water stored in the sump and the operation of the pump had the potential to mitigate, delay and/or dampen actual discharge salinity fluctuations. For example, with a large volume of saline water in the sump, a large reduction in salinity associated with fresh inflow would be 'hidden' by dilution of the saline water to some extent and be measured as a smaller reduction in overall pumped discharge salinity. If the sump storage was low at the time of inflow, the full magnitude of the reduced salinity could be measured after the pump was started. So the full range of discharge salinity fluctuations occurring in the drain channel is uncertain (Fig. 34).

When drain discharge was pumped frequently to keep sump water levels low the salinities represented the contributions by the whole drainage scheme. Groundwater inflow from the drain headwaters with salinities of around 20 000 mg/L would mix with downgradient water of 50 000 mg/L (Fig. 25) to deliver to the pump drain-flow with salinity of about 43 000 mg/L.

With delays in pumping water levels in the downstream end of the drain rose, reducing the more saline (50 000 mg/L) groundwater inflow from the surrounding landscape. With subsequent pumping salinity measurements would be skewed to the lower salinity water from the headwaters of the drain. This resulted in apparent reductions in both the average and flow-weighted salinities of the pumped and calculated drain discharges.

Longer-term salinity trends less affected by issues related to pump operation showed some but not a significant decline in drain-flow salinity beyond that caused by disconnecting the basin cut-off drain. While the drain-flow and basin leakage recirculation were combined the average flow-weighted salinity was 47 500 mg/L. From disconnecting the cut-off drain to the end of 2005, average flow-weighed salinity reduced to 43 000 mg/L, then to an average of 38 500 mg/L for 2006.

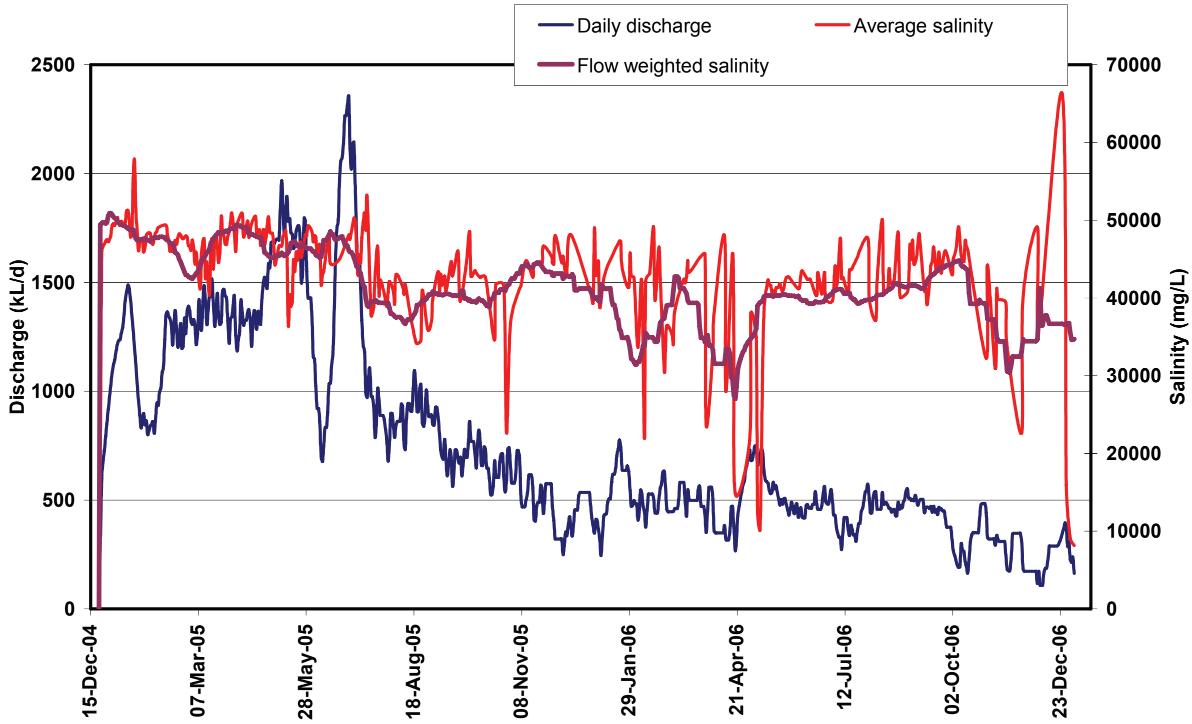

Figure 34 Main drain discharge average daily salinity and daily flow-weighted salinity

Discharge pH was mainly within the same range of values (6.5–8.0) as those of the surrounding watertable (Fig. 28). After the cut-off drain was disconnected in May 2005, the discharge pH tended to stabilise around 7.5–8.0, which were higher than the average values for the surrounding deep and shallow groundwaters (Fig. 35).

Like discharge salinity, the pH values were affected by the rate and timing of pumping from the drain and the contributing sources of groundwater inflow to the drain at the time. From the limited pH data available there appeared to be no consistent relationship between either the pumping rate or water level in the sump and pH. Lower pH values could be indicative of a rainfall contribution to drain-flow or that discharge was then dominated by groundwater inflow from the downstream end of the drain. The groundwater pH values at the downstream end of the drain in transects 4–5 were usually closer to the range 6.5–7.0 compared to the higher levels measured upstream (Appendix CD 4.2b).

There was no apparent trend in the discharge pH after disconnection of the cut-off drain (Fig. 35); the pH measurements were possibly too infrequent or irregular to identify any seasonal or short-term trends.

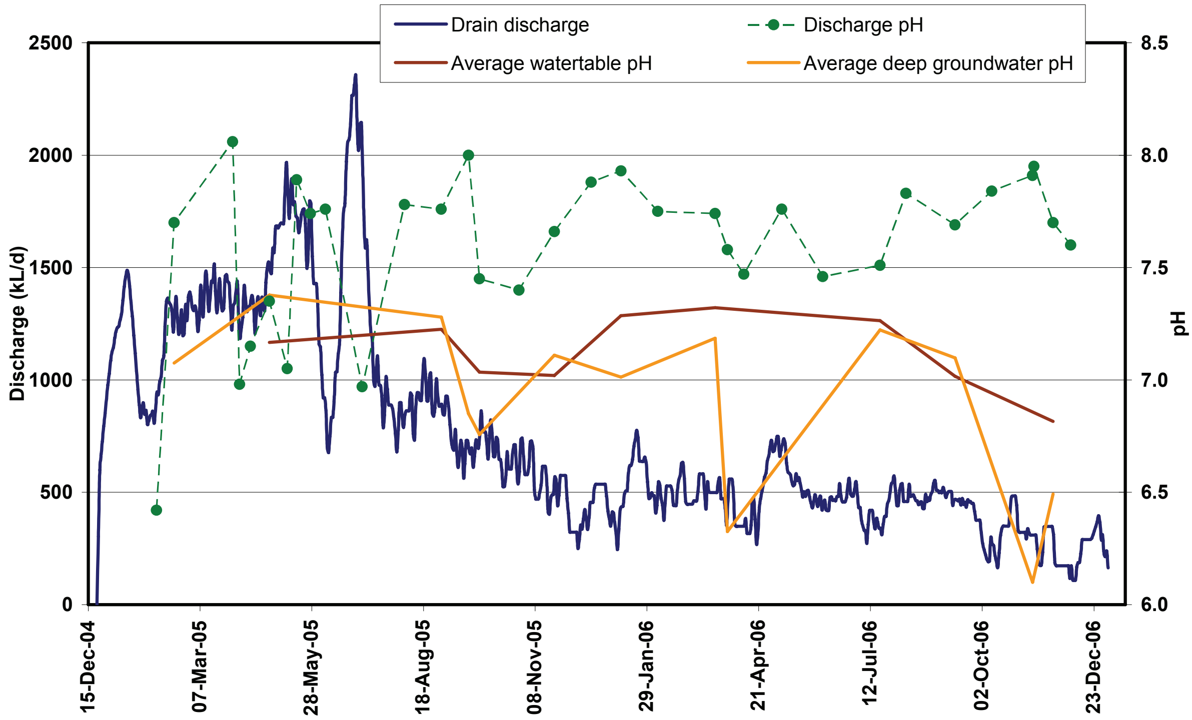

Figure 35 Main drain discharge and in-situ pH

Major ion concentrations measured in the drain were within the ranges for groundwaters. The discharge water quality was close to that of seawater except that carbonate levels were doubled. As discussed for groundwaters (Section 4.4), the neutral pH largely reduced the potential to dissolve and mobilise the heavy metals that are of concern in other Wheatbelt drainage schemes. The concentrations of heavy metals, total P and total N measured in the drain-flows were all within the ranges measured in the surrounding groundwaters (Section 4.4).

4.6 Mid-drain discharge

The mid-drain gauging station (025) was installed to measure and allow comparison of the proportions of the upstream to downstream discharges from the drain (Fig. 11). A total 20 100 kL discharge was measured over 45 days between 1 May and 15 June 2005, providing an average discharge 447 kL/d (Fig. 36). This equates to an average discharge rate of 0.97 L/s/km for the 5300 m of drain upstream of the mid-drain station. During the same period, the average pump discharge rate measured at the main-drain station (605) was equivalent to 2.47 L/s/km of drain (including basin leakage).

From 1 February to 31 December 2006 the average discharge rate from the mid-drain station had reduced to 0.36 L/s/km from the now relocated station with 5100 m of drain. In comparison, the average discharge at the main-drain station was 0.61 L/s/km from the total 8300 m of drain. During the 2006 period, the mid-drain station contributed 59% of the total drain discharge from 61% of the total drain length. The closeness of the proportional relationship between discharge and drain length suggests that groundwater inflow to the channel was fairly uniform along the drain length during this time.

This proportional relationship between the upstream and downstream lengths of drain and discharge appears to have been affected by the pumping from the drain. Allowing water to build up in the downstream end of the channel (Section 4.5) is thought to have reduced the proportion of discharge from the downstream drain section during the 2005 period.

During May–June 2005, the 20 100 kL discharge from the mid-drain station was 26% of the 76 900 kL total pump discharge measured at the main-drain station. After subtracting the estimated basin leakage recirculation rate of 476 kL/d (Section 4.5) the pumped discharge of drain-flow only was 55 500 kL. During this early period of drain-flow when there was a more concerted effort placed on pumping, discharge from the upstream station was only 36% of that of the downstream station. To generate the 35 400 kL of discharge between the mid-drain and main-drain stations, average drain-flow was equivalent to 3.37 L/s/km from the 2700 m of drain between the two stations.

There were too few mid-drain discharge measurements to demonstrate any relationship between discharge and rainfall (Fig. 36). The 2006 mid-drain discharge trend closely resembled that of the main drain, being relatively uniform and unresponsive (Fig. 33). Although there appeared to be some responses to daily rainfall these may in part have been due to rainfall contributions directly to the channel and drain-flow. In some cases, the magnitude of the responses tended to indicate the partial obstruction of the mid-drain gauging station control structure with debris. This was clearly the case in late October 2006 (Fig. 36) when the drain discharge response to the ~10 mm of rainfall greatly exceeded that expected for such an event.

Salinities reduced from 7 June 2005 to 21 January 2006 (Fig. 37) coinciding with disconnection of the cut-off from the main drain. Without recirculation of basin leakage water levels in the pump sump and main drain were lower, resulting in less frequent tail-water inundation of the mid-drain station.

The mid-drain station was not completely unaffected by tail-water until relocated 1200 m upstream from Holders Road (Fig. 11). At this site it measured what appeared to be lower salinity discharge from the upstream end of the drain. After relocation on the 21 January 2006 salinities fell as low as 5000 mg/L (Fig. 37), corresponding to salinities of some groundwaters measured in transects 1–2, upstream of this station (Appendix CD 4.2b).

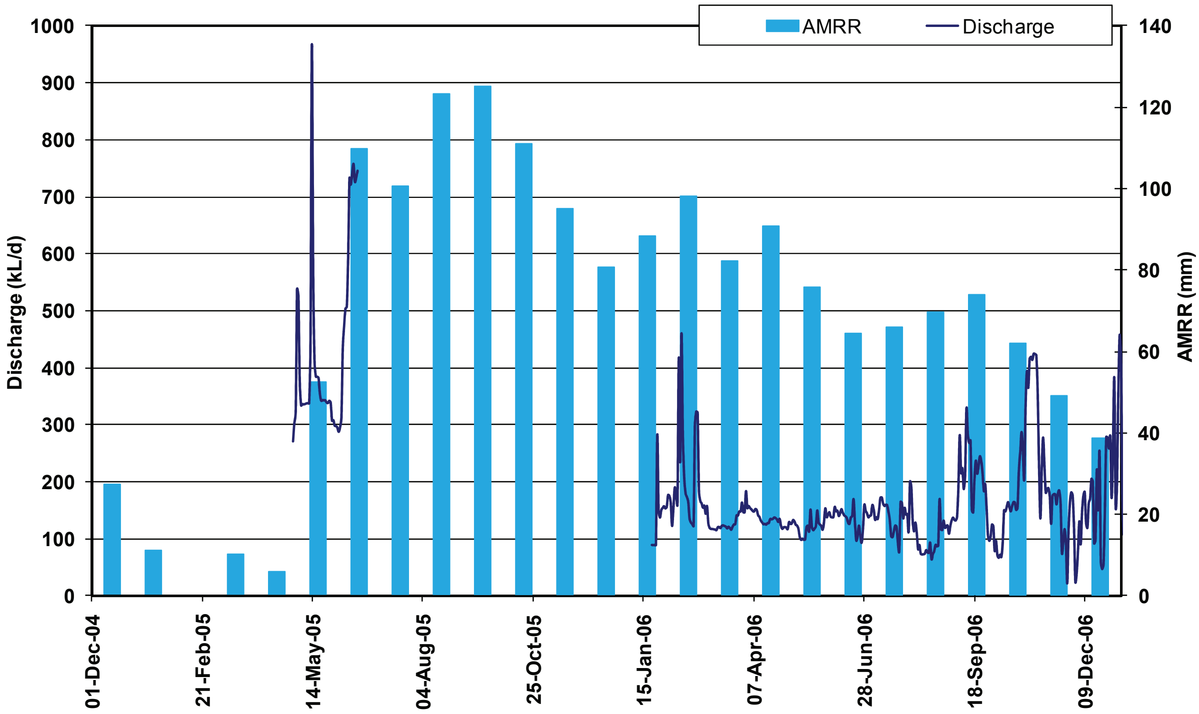

Figure 36 Discharge from the mid-drain station and AMMR trend

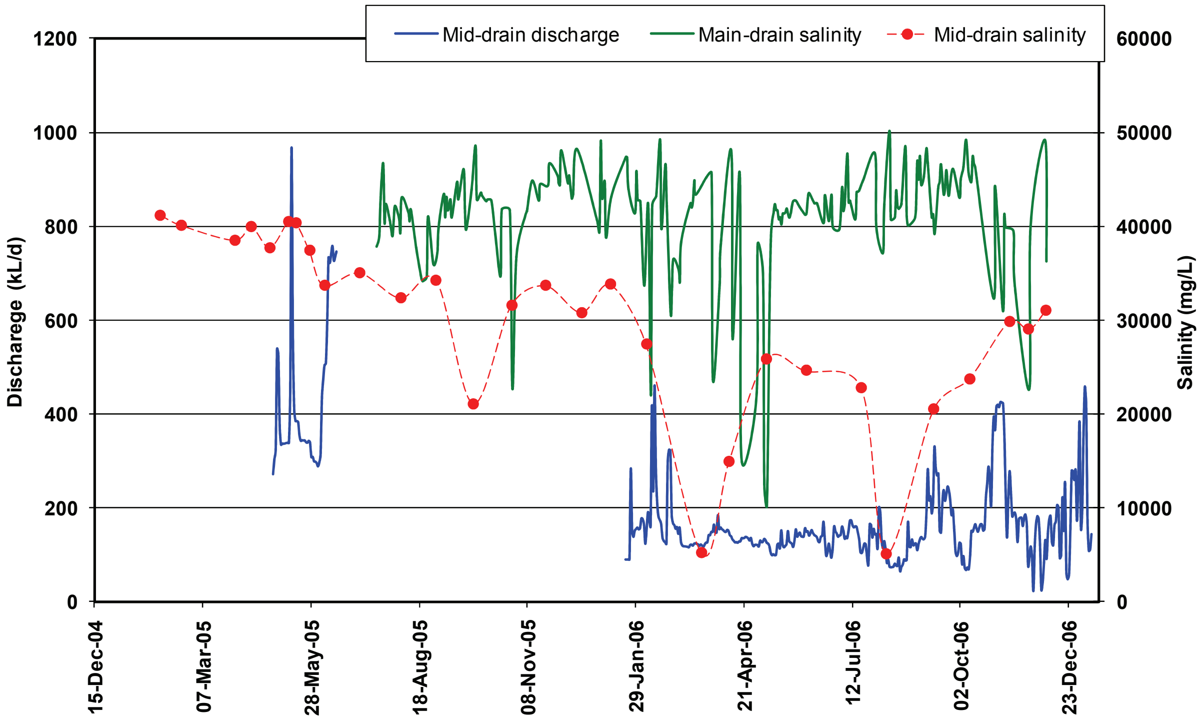

Figure 37 Salinity at the mid and main drain stations

The mid-drain discharge pH reflected an upward step trend, inverse to that of the discharge salinity (Fig. 38). These step-changes suggested the likelihood of discharge pH being affected by the same drain management responses as affected the salinities.

From 3 February to 7 June 2005, the average pH 7.5 discharge levels corresponded with those of the main-drain station (Fig. 38). From 7 June 2005, pH levels fluctuated within the average range of groundwater values for transects 1–2 (Figs 28 & 29) upstream of the mid-drain station. Towards the end of the record, discharge pH values rose to nearly 9.0, well above the levels measured in the surrounding groundwaters or the main drain.

Similar pH levels (close to 9.0) were only measured in the basin (Fig. 41), suggesting the possibility of interactions between the groundwater inflow to the drain and surrounding topsoil. Soil pH levels of 8.7 (in water) had been recorded at some sites upstream of the mid-drain station (Bell et al. 2009).

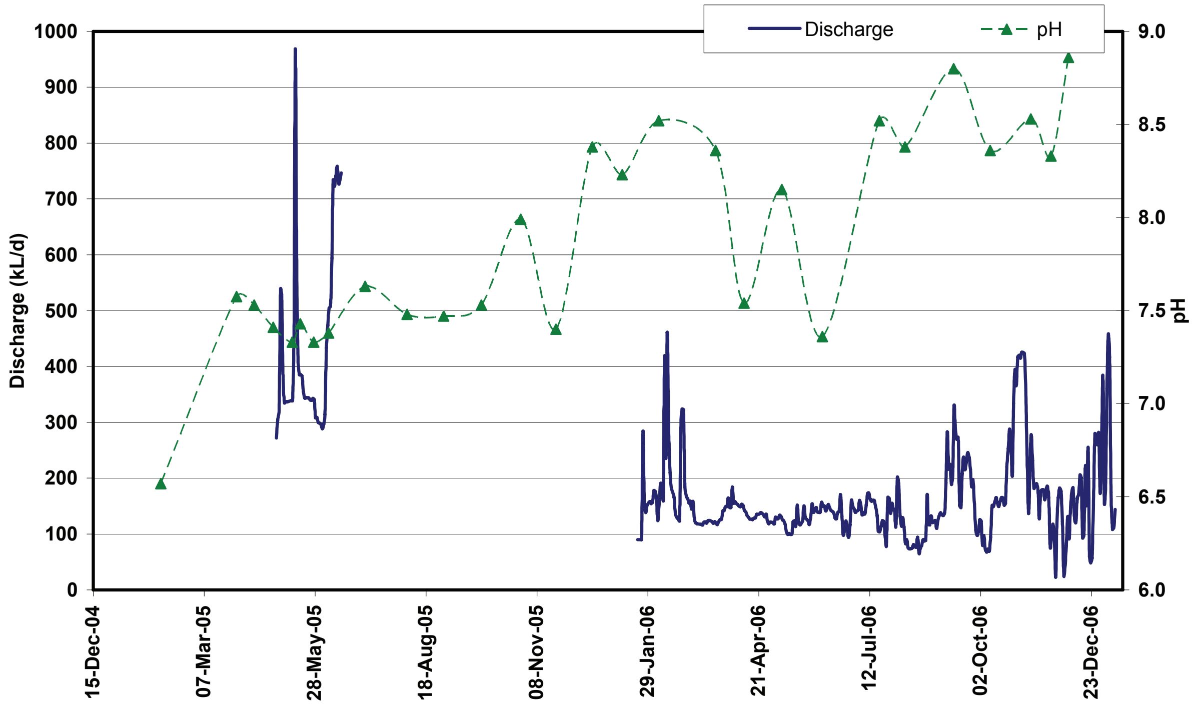

Figure 38 Mid-drain discharge and in-situ pH

4.7 Basin waters

The flatness of the basin site was well suited to constructing an evaporation basin. The maximum fall across the basin, from its SE corner to its NW side, was less than 0.2 m. Rising water could spread evenly across its entire floor area, maximising evaporative losses.

Drain discharge was first pumped into, and the water levels measured at, the NW corner of the basin. Measuring the filling began on 2 January 2005 from the elevation of 282.38 m AHD within the borrow area (Section 3.2). When the water level had risen by about 0.5 m it began to flood out of the borrow area and covered the natural ground surface in the western cell where the volume reached 25 000 kL by 20 March 2005 (Fig. 39) and began flowing into the eastern cell. From 20 March to 15 April 2005 flow between the cells was through a 150 mm PVC pipe through the dividing wall at an elevation of 283.70 m AHD (about 0.8 m above the natural ground). On 15 April 2005, the pipe was removed and the water flowed unrestricted between the cells at close to the original ground elevation of 282.90 m AHD.

On 10 June 2005 the basin volume approached its design capacity of nearly 50 000 kL. This prompted disconnection of the cut-off drain from the main drain on 15 June 2005 (Section 4.5) and construction of the 26.6-ha contingency evaporation basin (Section 3.1). The pump station was re-located to the SE corner of the eastern cell but the basin water level sensor was left at the NE corner of the basin.

On 6 July 2005 a 5 m wide section of the original basin's southern wall was breached to allow the water to flow into the contingency basin. The breach allowed the original basin to empty to approximately 15 200 kL (Fig. 39) with a water level elevation of 283.14 m AHD.

During this event 31 500 kL flowed from the original basin into the contingency basin. There was no further flow between the two basins after 11 July 2005. On 26 October 2005 the breach was repaired and the contingency basin was not used again during the monitoring period.

As the basin filled, the evaporative surface area in the western cell expanded rapidly as water inundated the entire floor area by 22 January 2005 (Fig. 39). Then there was a marginal daily increase in area as the water level rose until on 15 May 2005 the evaporative surface area again increased sharply as water flowed from the western cell into the eastern cell when the connecting pipe was removed and the dividing wall was breached.

The surface area peaked at the maximum of 49 000 m^2 on 6 July 2005, before stored water was released into the adjoining contingency basin (Fig. 39). Although the surface area was close to its maximum from 6 July to 12 November 2005, only about 10 000 kL of water was stored in the basin during this period.

When stored water in the basin falls below 6800 kL it recedes from the natural ground level into the borrow areas. In there the water has a maximum surface area of 15 300 m^2 which does not decrease much with further reductions in storage volumes and water levels. The volume can fall as low as 750 kL in each cell before the evaporative surface area starts shrinking significantly. From 12 November 2005 to the end of the project, the stored water in the basin was mostly within the borrow areas and its surface area rarely expanded above 15 300 m^2 (Fig. 39).

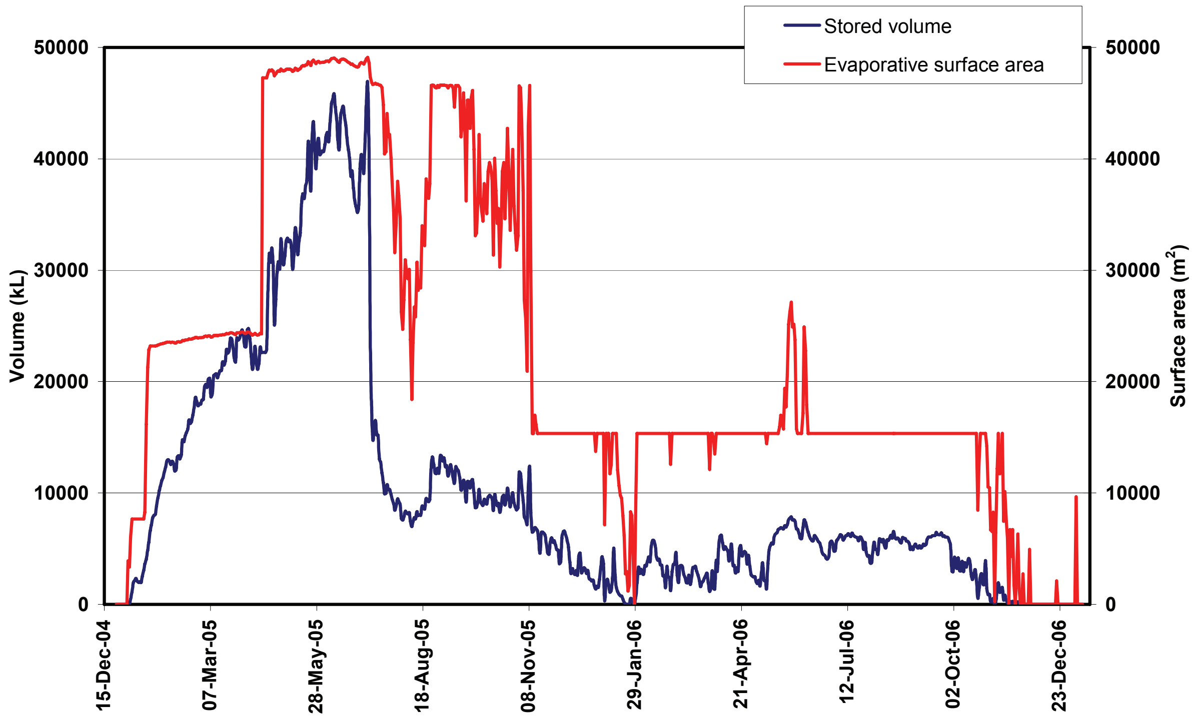

Figure 39 Basin storage and surface area

Basin water salinities rose significantly above those of the pumped inflow due to evapoconcentration only during October 2005–March 2006 (Fig. 40). During this time both water storage in the basin and the surrounding groundwaters were at their optimum levels to enhance evaporation and minimise leakage of the pumped inflow. Groundwater levels remained around 0.75 m below ground within 50 m of the basin at this time (Figs 21 & 22) while water levels in the basin were low, minimising the head difference and the associated potential for leakage. Nevertheless, basin storage was sufficient for much of the period to cover a large proportion of the basin area (Fig. 39).

Evaporation ranged from 6 to 18 mm/d during the October–March period (Fig. 2). An average evaporation rate of 11 mm/d was sufficient to evaporate the ~500 kL/d pumped drain discharge (Fig. 33) from a 4.5 ha evaporative area. Salinities began to rise again towards the end of the monitoring period in response to evapoconcentration, rising to around 7000 mg/L above the salinities of the pumped inflow.

High salinities in the basin at the commencement of pumping are consistent with both some evapoconcentration of salts and the recirculation of higher salinity groundwater from beneath the basin during February–May 2005. Groundwater salinities beneath the basin are about 50 000 mg/L (Figs 25 & 27). This saline water was captured and recirculated until diluted by rising rates of drain discharge from the end of April 2005.

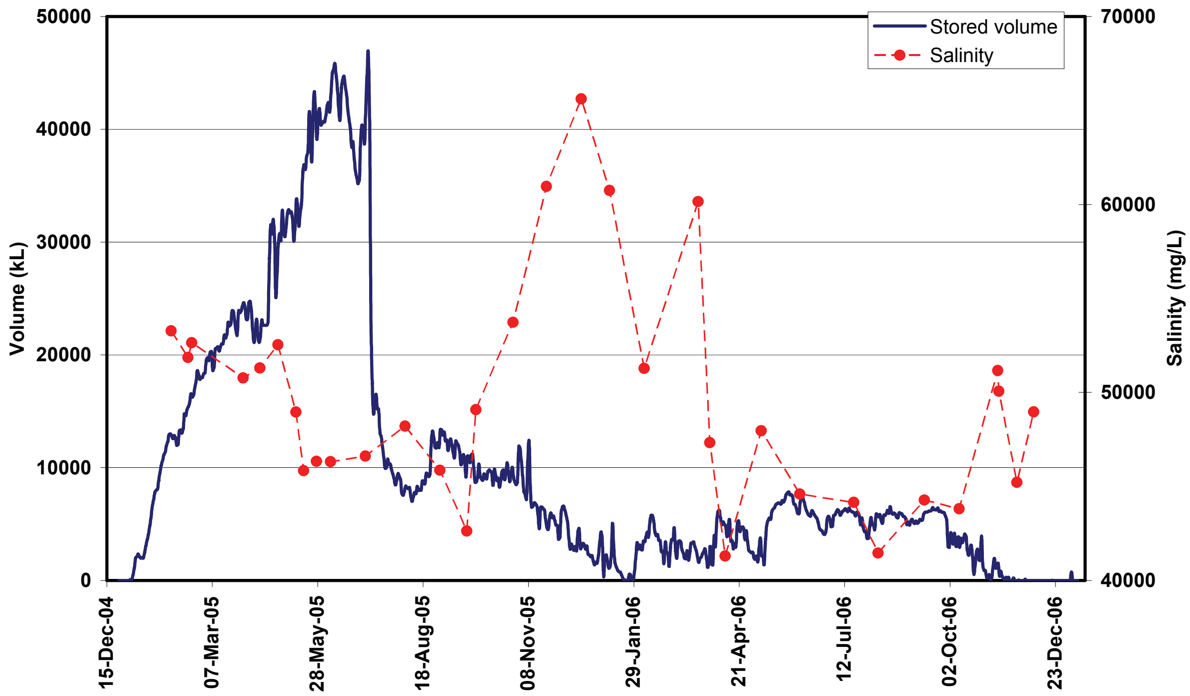

Figure 40 Basin water storage volume and salinity

The pH trend of the basin water closely corresponded to that of the drain discharge, after disconnection of the cut-off drain in June 2005 (Fig. 41). For the most part, the pH of the basin water remained 0.5 units higher than the drain discharge at around pH 8.25. As with the drain discharge (Section 4.5) there appeared to be no natural short- or long-term pH trends reflected in the measurements.

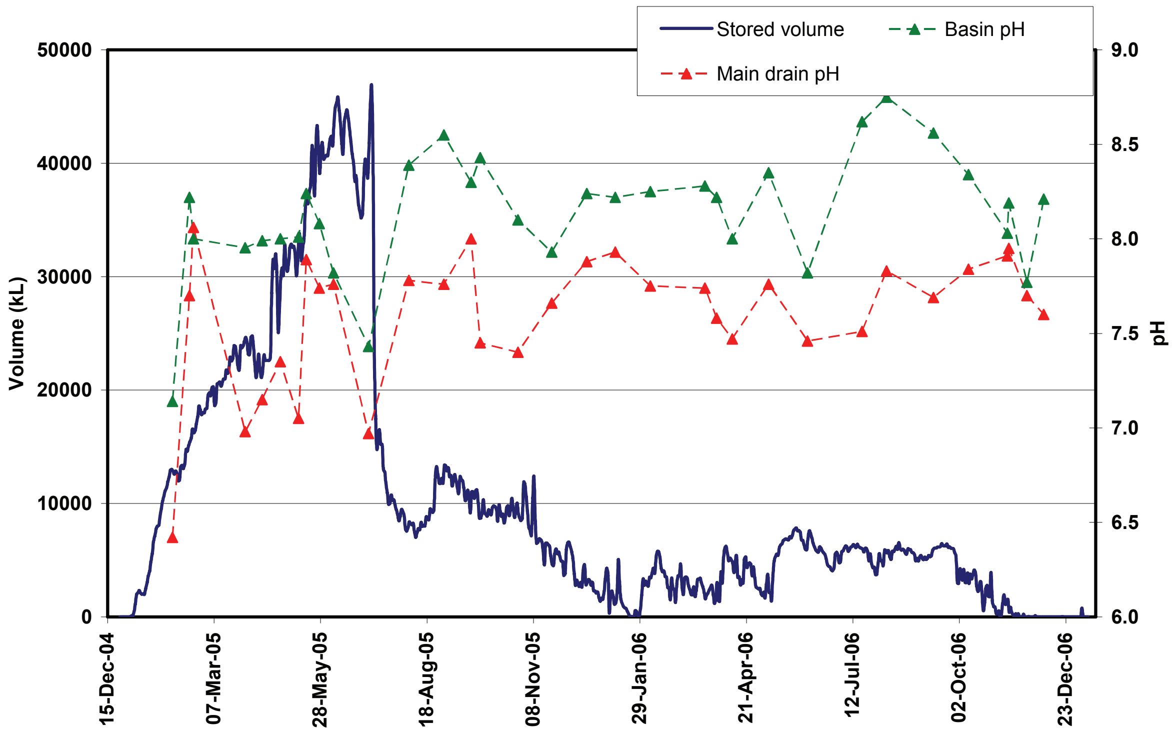

Figure 41 Basin water storage volume and pH and main drain discharge pH

The measured basin volume could be underestimated when it was less than 6570 kL, due to the basin's filling pattern and because the water level sensor was at the opposite end to the filling point. When the volume was greater than 6570 kL both borrow areas were full, the water levels above the 282.90 m AHD elevation of the divide between them and the water levels in each cell rose and fell in unison. When the volume fell below 6570 kL and the 282.90 m elevation, the water levels in the borrow areas could fluctuate independently (Fig. 42).

The significance of this filling pattern becomes apparent when comparing the rate of pumping into the basin with the rate of filling from initial low storage levels. The water level in the eastern cell had to rise above 282.90 m to allow water to flow into the western cell and be recorded as a water level rise for the whole basin. The assumption was that as long as the basin water level was rising, the eastern cell water level must have been at or above 282.90 m elevation and its volume at 3100 kL to enable the water to flow from the western to the eastern cells.

After the project was completed local farmer groups experimented with growing seaweed in the basin and newly constructed ponds at its western end (Fig. 42).

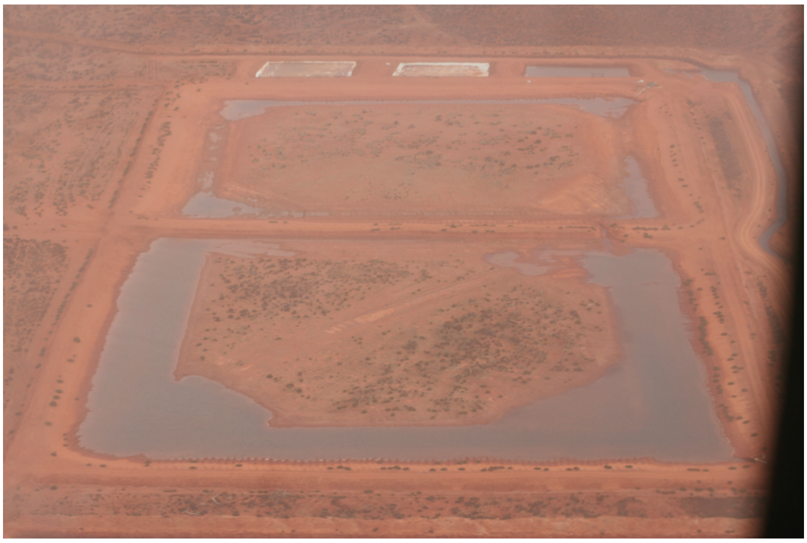

Figure 42 Partially filled basin viewed from the east with seaweed ponds in the background

5 Water and salt balances

5.1 Drainage and evaporative basin sites

The Morawa drain was constructed in a 1–2 km-wide valley floor with an underlying 30–40 m thick semi-confined and unconfined aquifer. The valley flanks extend away from the valley floor for up to 5 km at gradients generally less than 2%. The regolith beneath the valley flanks is thought to be up to 30 m thick and essentially provides for the continuation of the valley floor aquifer.

The groundwater depth beneath the valley flanks varies but reaches elevations well above the valley floor watertable. The longitudinal gradient along the valley floor is low, resulting in sluggish surface movement and effectively no groundwater movement. In-situ and runoff recharge combine with aquifer discharge from the hillsides to cause groundwater to accumulate and rise beneath the valley floor.

Over the recent past the groundwater levels beneath the valley floor appear to have been at an equilibrium level of about a metre below the ground surface. This equilibrium is between the inputs to the valley floor aquifer from recharge and hillside aquifer discharge, and capillary discharge from the watertable and surface evapotranspiration. Capillary discharge and surface evaporation deliver and accumulate salts from groundwater onto or near the land surface.

Under these conditions the valley floor and its aquifer are operating as the groundwater outlet for the surrounding catchment through groundwater discharged by evaporation and, on rare occasions, seepage. The view formulated in this report is that constructing the groundwater drain in the valley floor enhanced discharge from the aquifer. However, it is more complex to demonstrate why, despite discharge being greatly enhanced by the drain, the adjacent watertable did not fall significantly as expected.

The most concise method of exploring all possible effects of the drain was through pre- and post-drain water balances for the drain and surrounding aquifer. The groundwater components of the drainage site water balance provided by Chandler & Coles (2003) to solve the water balances for Morawa using available data were complex. The water balance:

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

used recharge (R) combined with shallow seepage in and out of the site ($S_{IN} - S_{OUT}$) and upward and downward groundwater percolation into an underlying aquifer ($Z_{IN} - Z_{OUT}$) to produce groundwater inflow to the drain (Q). For Morawa, complex groundwater modelling supported by further investigatory drilling would have been needed to solve for the shallow seepage and deeper percolation components of this water balance equation.

Given their apparent connectivity beneath the Morawa drain, complex groundwater modelling to isolate shallow and deep groundwater systems was considered unnecessary. However, because this was a ‘single’ drain within an extensive groundwater system, the need to account for the lateral movement of groundwater into and out of the drainage site remained.

To achieve this without complex modelling involved placing an arbitrary ‘porous’ groundwater boundary around the aquifer adjacent to the drain. The movement of groundwater through this boundary then becomes part of the water balance. This is subsequently referred to as aquifer discharge (A_D), into or out of the drainage and basin sites (Appendix F).

Because one objective was to explore the relationship between changing water balance and groundwater levels, the distance between the boundary and drain needed to realistically reflect the distance within which the drain was expected to affect the groundwater and watertable height. At Morawa, most of the watertable responses extended 275–400 m from the drain and basin (Section 4.1). The relationship between the watertable responses at increasing distances from the drain and groundwater inflow to the channel was further explored in an attempt to confirm the response relationship between watertable change and drain discharge (Appendix C).

In Figure 43 the groundwater boundary to the drain is 400 m from the drain centre line, capturing the watertable responses. This equates to the outer extents of the transect 1–4 bores which are the main sources of reliable groundwater information. The area encompassed within the boundary is referred to as ‘the drainage site’, and has an area of 597 ha and a perimeter of 15 200 m.

The same groundwater boundary approach delineates ‘the evaporation basin’ site (Fig. 43). The boundary at 400 m from the basin perimeter encompasses the area within which watertables appear to have responded to basin leakage. The basin site encompasses 61 ha in addition to the basin area, and has a boundary perimeter of 2600 m.

The movement of groundwater between the basin and drain sites does not form a significant component of the water balances after disconnection of the cut-off drain. So the 764 m of common boundary between the basin and drain sites has not been included in the above mentioned perimeter lengths used to calculate aquifer discharges (Appendix F).

The 61 ha basin water balance site does not incorporate the contingency basin. The water balance of this basin was not assessed due to its short-term use when water was released into it from the main basin on 6 July 2005. The release of the 31 900 kL of water and associated salt are part of the water and salt balances of the main basin (Section 5.3).

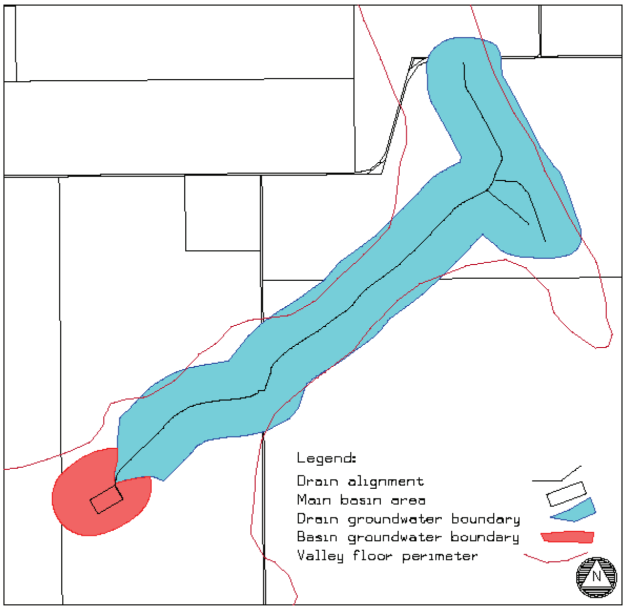

Figure 43 Representation of drainage and basin layout and associated sites for water balance calculations

5.2 Water and salt balances for the drain

Drain site water balance

The water balance of the Morawa drainage site shows that, even after drainage, the potential loss of groundwater by capillary rise and evaporation was several times greater than that removed by the drain. The water balances of the drain and drainage site, although interlinked, were calculated separately to distinguish the individual effects on these of the same naturally occurring variables.

The water balance of the drain itself captures the inputs of groundwater inflow from the surrounding aquifer, intermixed with sporadic rainfall into the channel and runoff from the drain structure (Fig. 44). Outputs from the drain are its measured discharge (Section 4.5) and the evaporative loss of drain-flow directly from the channel.

Groundwater inflow to the drain was from groundwater outflow from its surrounding aquifer beneath the drainage site. Groundwater outflow–inflow is the only component linking the water balances of both the drain and drainage site. Groundwater outflow and groundwater evaporation are the two outputs from the drainage site water balance that reflect the loss of groundwater from aquifer storage and potential for lowering of the watertable (Fig. 44).

Rain-fed recharge and inward flowing aquifer discharge both contributed to aquifer storage and potential watertable rise beneath the drainage site. While aquifer discharge is reflected directly in the water balance equation below, recharge is embedded in the relationship between precipitation (rainfall) and the pre-groundwater recharge losses of evapotranspiration, runoff and change in soil water storage.

Pre-groundwater losses represent rain water lost from the drainage site before the rest had an opportunity to become recharge. There was no measurement of or attempt to differentiate between the three loss components, a value for which was only provided to contextualise the other values in the water balance, and balance the equation. Groundwater recharge is the difference between precipitation and the pre-groundwater losses in the water balance. Although groundwater recharge was calculated (Appendix E) it has not been expressed directly in Equation 1.

With reference to Figure 44 the water balance of the drainage site was calculated with the equation:

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

Where: P : Precipitation (Section 2.1)

A_D : Aquifer discharge (Appendix F)

$(ET + RO + \Delta S_S)$: Evapotranspiration, runoff and change in soil water

A_S : Aquifer storage change (Appendix D)

G_E : Groundwater evaporative loss potential (Appendix G)

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

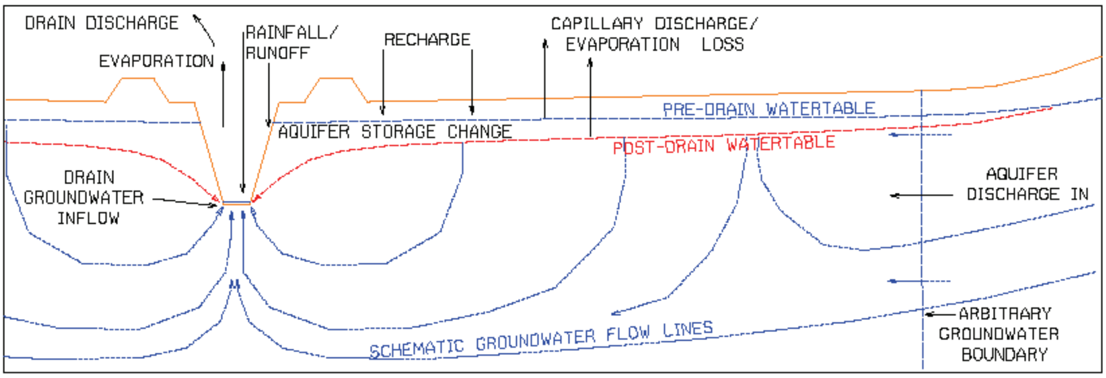

Figure 44 Conceptual water balance of the drainage site and evaporation basin site

Water balance results are often expressed in terms of the total volumes of water relating to individual components, millimetres depth equivalent across the area of the drainage site or as a percentage of rainfall. Substituting the measured and calculated volumes (see Appendix CD 5.9) into the water balance equation (Equation 1) for the Morawa drainage site from 13 July 2004 to 31 December 2006 produced:

3 038 700 + 513 700 = 1 690 400 + (-28 200) + 1 328 400+ 561 800 (kL)

There was no way of making any true annual comparisons between the pre- and post-drain results because of the short pre-drain monitoring period and the significant influences of seasonal variability. The most important result of pre- and post-drain comparisons was that it showed that before drainage the only pathway for groundwater loss was evaporation following capillary rise, as reflected by the zero value for the pre-drain groundwater outflow (Table 2).

The construction of the drain introduced groundwater outflow into the landscape and equation. Although the relative contributions of groundwater to this outflow are unknown, without the drain as an outlet, the 561 700 kL of groundwater outflow, the equivalent of 23% of 2005–06 rainfall, would have remained in storage or been lost by groundwater evaporation.

During the pre-drain period 4.9 mm of rainfall went into aquifer storage reflected in the watertable rise evident in many of the pre-drain groundwater hydrographs (Section 4.1). Post-drainage aquifer storage decreased by 9.6 mm following an overall decline in watertables across the drainage site.

Aquifer storage change is a dynamic expression of groundwater moving into and out of the aquifer, and not a contributor to the volume of water involved in the overall water balance. Only if aquifer storage changes between the start and end of the water balance simulations does it need to be accounted for in the overall result. Hence the timing of the start and end of the water balance calculations will affect the aquifer storage change value.

Table 2 Pre- and post-drain water balances expressed as mm depth across the drainage site

	P	**A_D**	**$(ET+RO+\Delta S_S)$**	ΔA_S	**G_E**	**G_O**
Total (20/5/04–13/12/06)	509.0	86.0	283.1	-4.7	222.5	94.1
Pre-drain (20/5/04–22/12/04)	101.5	14.1	66.5	4.9	44.3	0.0
Post-drain (22/12/04–31/12/06)	407.4	71.9	216.7	-9.6	178.2	94.1

In practice, groundwater evaporation is a component of the other soil water losses but has been isolated in this water balance because it directly affects watertable levels and is an indicator of drainage success. The calculated potential groundwater evaporative losses are likely to be far higher than actual losses across the drainage site (Table 2). The potential evaporative losses were calculated (Appendix G) firstly, to demonstrate how much groundwater was involved, and secondly as a means of exploring changes in the capillary discharge of salts in the subsequent salt balances.

A more realistic value for groundwater evaporative loss was estimated by restricting the water balance equation to only groundwater components, creating the groundwater balance equation (Equation 2). This was done by directly introducing recharge (U) (Appendix E) in combination with aquifer discharge from Equation 1, as the input values for Equation 2.

Recharge and aquifer discharge become the two direct sources of groundwater to the drainage site in Equation 2 (Fig. 44).

Groundwater outflow to the drain and evaporation by capillary rise are the two sources of potential groundwater loss from the drainage site (Fig. 44). Change in aquifer storage is a reflection of groundwater movement from the recharge to discharge sides of the water balance equation and can sit on either side of Equation 2 providing it is given the appropriate negative or positive sign. In Equation 2, change in aquifer storage has been retained as a discharge, as in Equation 1.

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

The substitution of the measured and calculated values into Equation 2 substantially reduces groundwater evaporation, particularly for the post-drain condition (Table 3).

There were approximately six months of pre-drain monitoring and 24 months post-drain monitoring. To express both sets of results on an approximate annualised basis the pre-drain results substituted into Equation 2 were multiplied by two and the post-drain results divided by two (Table 3).

In the pre-drain equation, 53 mm/yr of combined recharge and aquifer discharge produced 43 mm/yr of groundwater evaporation, with the remaining 10 mm/yr going into aquifer storage. For the post-drain period, 47 mm/yr of the 57 mm/yr of recharge and aquifer discharge became groundwater outflow and an additional 15 mm/yr was lost by groundwater evaporation. This imbalance between recharges (57 mm/yr) and discharges (62 mm/yr) was corrected by the discharge of 5 mm/yr of groundwater released from aquifer storage. The two options available for this loss of 5 mm/yr of groundwater from aquifer storage were via groundwater evaporation and/or groundwater outflow.

Table 3 Pre and post-drain groundwater balances expressed as mm depth across the site

	U	$\mathbf{A_D}$	ΔA_S	$\mathbf{G_E}$	$\mathbf{G_O}$
Total (20/5/04–13/12/06)	54.0	86.0	–4.7	50.7	94.1
Pre-drain (20/5/04–22/12/04)	12.9	14.1	4.9	21.4	0.0
Pre-drain (annualised)	25	28	10	43	0
Post-drain (22/12/04–31/12/06)	41.9	71.9	–9.6	29.3	94.1
Post-drain (annualised)	21	36	-5	15	47

Drain channel water balance

Groundwater outflow from the aquifer became groundwater inflow to the drain. In the channel, groundwater inflow intermittently mixed with rainfall and runoff and what did not evaporate became drain discharge (Fig. 44). Before the cut-off drain was disconnected from the main drain, leakage from the basin combined with the drain discharge before being pumped into the basin.

The water balance of the drain from inflow to discharge is expressed as:

$$Q + E = G_I + (RF + RO)$$ *Equation 3*

Where: Q: Discharge from the drain outlet

E: Evaporation from the drain channel

G_I: Groundwater inflow to the channel from groundwater outflow

(RF + RO): Rainfall and runoff accessions to drain-flow

During the period from drain construction to December 2006, the water balance in the channel was:

$$501\,800 + 65\,100 = 561\,800 + 5100 \text{ (kL)}$$

Note the relationship between the 561 800 kL of groundwater inflow in Equation 3 to groundwater outflow in Equation 1. Overall, the volume of groundwater inflow to the drain exceeded the total volume of drain discharge because 65 100 kL of in-channel water evaporated (Appendix B). Even with the addition of rainfall, monthly drain discharge never exceeded groundwater inflow, with evaporative loss being the difference between the drain discharge and groundwater inflow hydrographs (Fig. 45).

The volume of water pumped into the basin included drain discharge and recirculated basin leakage until 15 June 2005. The final volume of water being lifted into the basin is expressed as:

$$P_Q = Q + R$$ *Equation 4*

Where: P_Q: The volume pumped from the drain sump into the basin (Section 4.5)

Q: From drain discharge Q, above

R: Recirculation or leakage water from the basin (Appendix B)

Substituting the values into Equation 4 results in:

$$549\,400 = 501\,800 + 47\,600 \text{ (kL)}$$

Pumped discharge only exceeded drain discharge until the cut-off drain was disconnected in June 2005 after which pump discharge directly equated with drain discharge (Fig. 45). Before the disconnection, basin leakage recirculation increased to nearly 12 000 kL in May 2005, making up 23% of the month's pumped discharge.

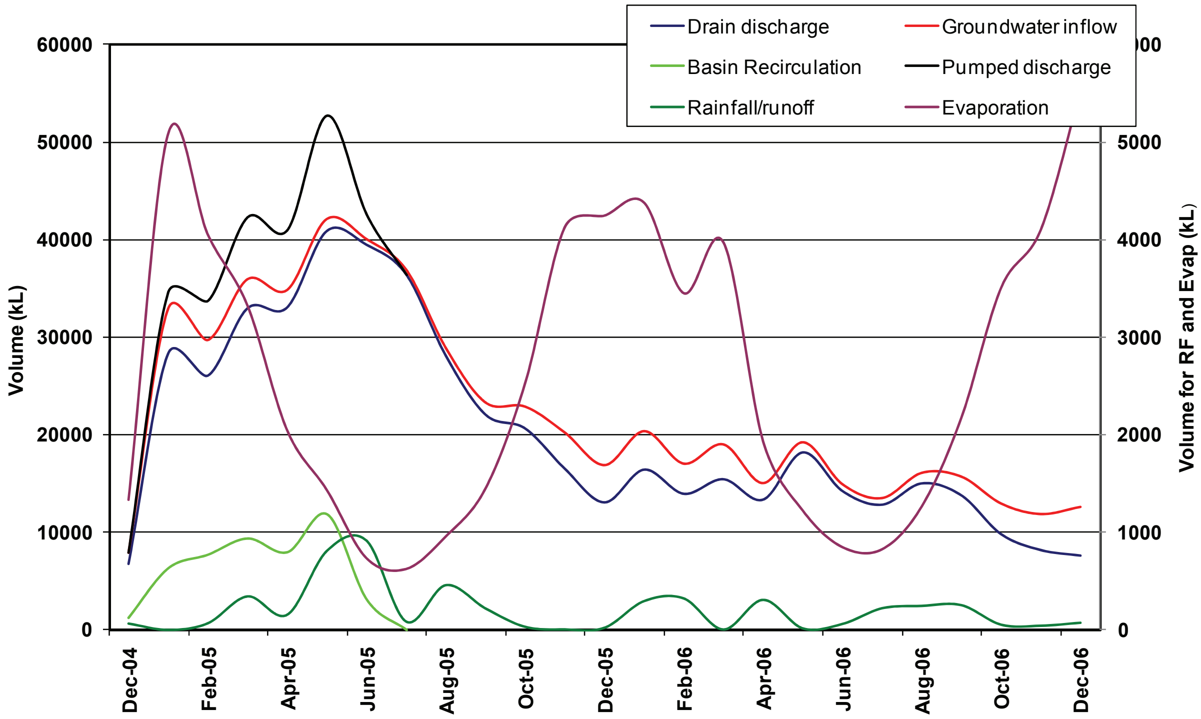

Figure 45 Monthly in-channel water balance for the Morawa drain

Drainage site salt balance

Dissolved in the 549 400 kL of water pumped into the basin were 23 600 t of salts (Section 4.5). Before the cut-off drain was disconnected from the main drain, 2800 t was recirculated between basin leakage and drain (Appendix B) and the remaining 20 800 t in the drain discharge was transported into the channel by groundwater inflow.

Recirculated basin leakage was responsible for an average 3400 mg/L increase in the salinity of the pumped drain discharge from 22 December 2004 to 15 June 2005. The average salinity of the basin leakage into the drain was 59 500 mg/L: higher than the salinity of the stored basin water during this time (Fig. 40) but closer to the salinity of the groundwater beneath the basin (Appendix CD 4.2b).

The salt balance of the drainage site cannot be solved because the mass of salt exported by capillary rise and subsequent surface runoff is unknown (Table 4). The salt-balance equation below contains the same variables as Equation 1 (drainage site water balance). In the salt-balance equation the groundwater variables have been assigned a salinity of 37 314 mg/L (Appendix B). This represents the average groundwater inflow salinity to the drain from the surrounding aquifer and was used to convert the volumes of saline water to salt loads. The mass of salt falling onto the site in rain was not measured but, unlikely to have contributed significantly to this short-duration salt balance, has been assumed to be zero.

The salt balance of Equation 1 for 22 December 2004–12 December 2006 produced an apparent imbalance on the loss side of about 50 100 t:

$$0 + 19\,200 = ? + (-1100) + 49\,600 + 20\,800 \text{ (t)}$$

The only measured value in the equation is from groundwater inflow. Values for aquifer discharge and storage change may vary considerably dependent on actual salinities and, for aquifer discharge, rates of groundwater movement. Although evaporation of groundwater by capillary discharge had the potential to transport and deposit 49 600 t of salt near the soil surface, the proportions of this salt removed from the drainage site in surface runoff or leached back into the soil are unknown.

Similarly, it is not known if the salt load from the post-drain aquifer storage reduction was removed from the drainage site (Table 3). If the storage change was in response to the drain, the salt should have drained away with the water but if the storage change was driven by evaporation the salt would have remained in the aquifer or been transported to the soil surface by capillary rise.

Putting the salt balance values into the groundwater balance Equation 2 produces more realistic results for salt deposition onto the soil surface by capillary discharge than Equation 1. In the approximately two and a half year salt balance simulation (Appendix G), an estimated 11 300 t of salt was transported to the soil surface by capillary rise and evaporation. If distributed evenly across the 597 ha drainage site (Section 5.1) and expressed as an average annual amount, this mass represents a 7.6 t/ha/yr average annual deposition rate (Table 4).

The most notable differences between the pre- and post-drain periods are the post-drain reduction in groundwater evaporation-driven salt deposition, and, of course, the salt exported from the site in groundwater outflow. As for the results of Equation 1, the mass of salt removed from the site in surface runoff is unknown as is the mechanism of salt loss from the reduction in aquifer storage.

Table 4 Representative annualised pre- and post-drain site salt balances in t/ha/yr

	U	A_D	ΔA_S	G_E	G_O
Total t/ha/yr	0	12.8	-0.7	7.6	13.9
Pre-drain t/ha/yr	0	10.5	3.7	15.9	0
Post-drain t/ha/yr	0	13.4	-1.8	5.5	17.4

After removing the unknowns the variables left enabling comparison between the pre- and post-drain conditions are aquifer discharge and groundwater outflow, giving in the equation, if balanced:

$$A_D = G_O \qquad \textit{Equation 5}$$

For the pre-drain period there was an imbalance in the equation as 10.5 t/ha/yr of salt was transported to the drainage site in aquifer discharge (Table 4). It was unknown if any of this salt was subsequently removed by groundwater discharge and runoff. In the post-drain period there was also an imbalance in the equation as 13.4 t/ha/yr of salt was transported to the drainage site and 17.4 t/ha/yr was removed by the drain, a net salt export of 4 t/ha/yr by the drain.

5.3 Water and salt balances for the basin

Water balance for the basin

The water balance of the evaporation basin and 60.5 ha basin site revealed that 79% of water pumped into the basin was lost to leakage which mostly evaporated within 400 m of the basin. Only about 12 500 kL of groundwater was lost from the basin site. Leakage, evaporation and other elements of the water and salt balances that affect the basin and basin site are illustrated in the conceptualised water balance (Fig. 46) and captured in Equations 6 and 7.

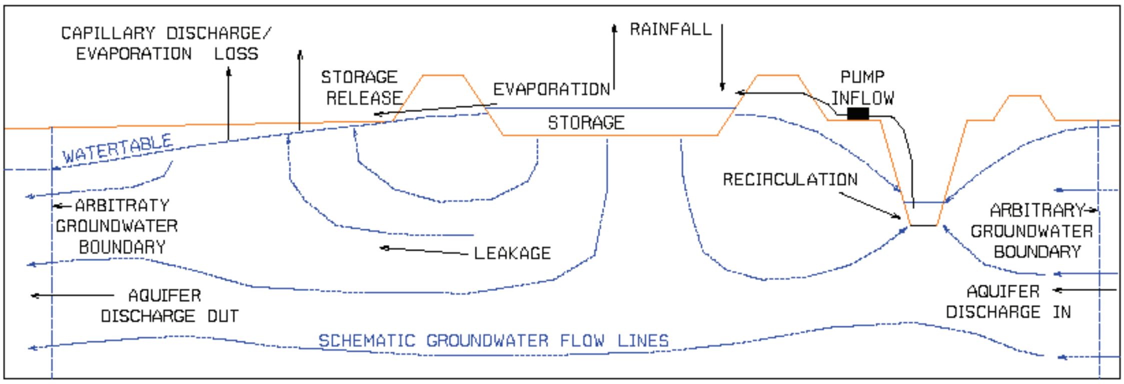

Figure 46 Conceptualised basin water balance

The basin water balance equation is:

$$P_Q + P = E + D + L + S \qquad \textit{Equation 6}$$

Where: P_Q : The volume pumped into the basin from the sump and drain (Section 4.5)

P : Precipitation to the basin surface area (Section 2.1)

E : Evaporation loss from the basin (Section 2.1)

D : Water released from the basin to contingency basin (Section 4.7)

L : Leakage

S : Volume of water in storage on 31 December 2006 (Section 4.7)

From the start of pumping on 22 December 2004 to 31 December 2006, the basin water balance was:

$$549\,400 + 20\,600 = 101\,800 + 31\,900 + 433\,200 + 3100 \text{ (kL)}$$

The basin and drain water balances are linked by the 549 400 kL discharge from the drain (Equation 4) pumped directly into the basin (Equation 6). The pumped discharge and inflow hydrographs from the drain and basin are directly comparable (Figs 45 & 47). The main sources of basin loss were leakage at 78.9% of pumped inflow, and evaporation, 18.5% (Appendix CD 5.8).

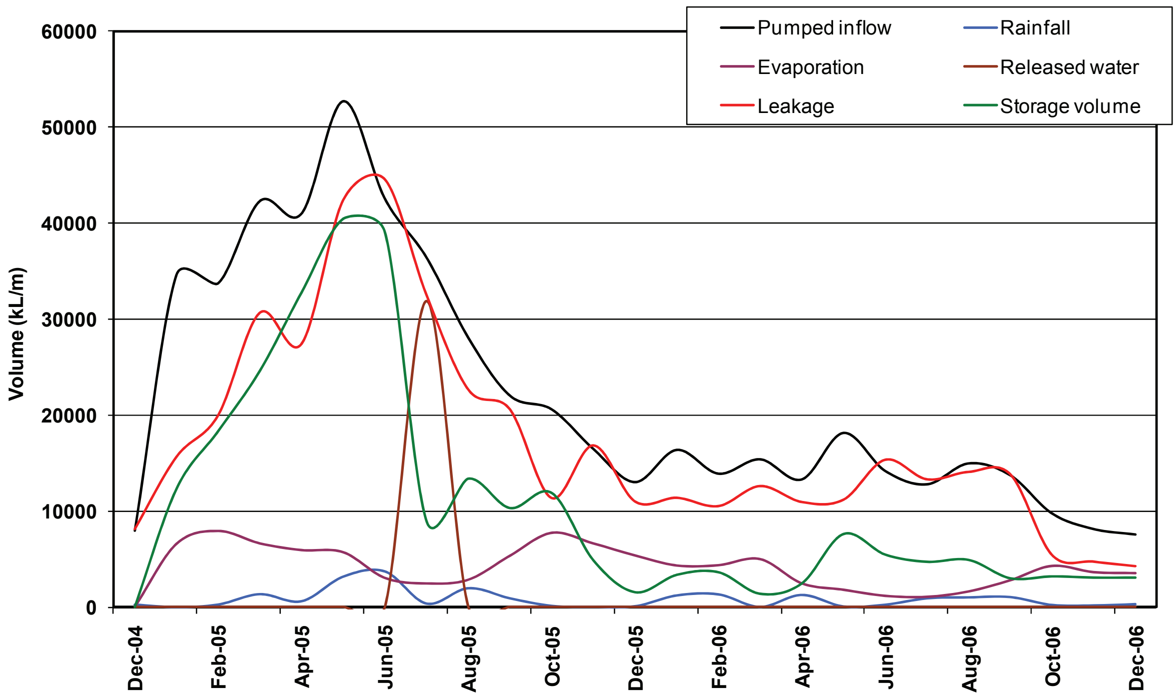

Figure 47 Basin monthly water balances

Water balance of the basin site

Leakage was the most significant input to the water balance of the 61 ha basin site as expressed within the water balance equation for the site:

$$P + A_{D\,IN} + L = (ET + RO + \Delta S_S) + \Delta A_S + G_E + A_{D\,OUT} + G_O \qquad \textit{Equation 7}$$

Where: P: Precipitation

$A_{D\,IN}$: Aquifer discharge groundwater flow into the site

L : Basin leakage from above

($ET + RO + \Delta S_S$): evapotranspiration, runoff and change in soil water

ΔA_S : Change in aquifer storage

G_E : groundwater evaporative loss by capillary discharge

$\mathbf{A_{D\,OUT}}$: Aquifer discharge groundwater flow out of the site

$\mathbf{G_O}$: Groundwater outflow from the site into the cut-off drain

The values for the basin site are for the entire monitoring period 20 May 2004 to 31 December 2006. Basin leakage is the largest component of the site water balance and provides the link between Equation 6 for the basin and Equation 7 for the site.

308 200 + 7600 + 433 200 = 480 100 + 22 700 + 186 100 + 12 500 + 47 600 (kL)

Before water was pumped into the basin, the pre-drain water balance of the basin and drain sites were similar (Table 2) with most of the rainfall lost as evapotranspiration and runoff, and groundwater lost by capillary discharge (Table 5). Aquifer discharge into the basin site from surrounding elevated land was similar to that of the drain site and aquifer discharge from the pre-basin site was practically zero.

During the pre-basin period no significant aquifer discharge out of the basin site was measured (or expected) because the groundwater levels within the site were at equilibrium with those outside. This presented the condition where no hydraulic gradient existed between the two to drive groundwater flow out of the site.

Pumping into the basin and subsequent leakage produced a head difference between the watertables inside and outside the basin site as a groundwater mound developed beneath the basin (Figs 20–22). The head difference drove 12 500 kL of aquifer discharge from the site (Equation 7) into the down slope groundwater system. This was equivalent to 20.7 mm across the site (Table 5).

The higher watertable within the basin site significantly increased the potential loss of groundwater by evaporation after seepage and capillary rise (Table 5). Figure 48 shows leakage water outside the SE wall of the basin indicating that the watertable is at and above the land surface. Because this water did not run off the site or into the cut-off drain it was not considered as groundwater outflow.

Table 5 Basin site water balances expressed as mm for each period

	P	$A_{D\ IN}$	L	$(ET+RO+\Delta S_S)$	ΔA_S	G_E	$A_{D\ OUT}$	G_O
Total (20/5/04–13/12/06)	509.0	12.5	715.5	792.9	37.4	638.0	20.7	78.6
Pre-basin (20/5/04–22/12/04)	101.5	6.6	0.0	92.7	6.4	10.7	0.0	0.0
Post-basin (22/12/04–31/12/06)	407.4	5.9	715.5	700.2	31.0	627.3	20.7	78.6

Figure 48 Leakage outside the SE basin wall

Restricting the water balance equation of the basin site to reflect only groundwater components allows recharge (U) to be in the equation and more realistic values for groundwater evaporation to be calculated. The restricted groundwater balance equation for the basin site becomes:

$$A_{D\ IN} + L + U = + \Delta A_S + G_E + A_{D\ OUT} + G_O \qquad \textit{Equation 8}$$

Groundwater evaporation was the largest source of groundwater loss from the basin site (Table 6). Converted to annual values, the evaporation was nearly 15 times the pre-drainage groundwater evaporation rate and accounted for nearly 90% of the basin leakage. The reduced post-basin aquifer discharge into, increased aquifer discharge from the site and the increased aquifer storage confirmed that groundwater levels beneath the site rose in response to the basin leakage.

Table 6 Representative annualised basin site groundwater balances expressed in mm/yr

	$A_{D\ IN}$	L	U	ΔA_S	G_E	$A_{D\ OUT}$	G_O
Total mm/yr	5.0	286.2	18.7	15.0	255.2	8.3	31.4
Pre-basin mm/yr	13.2	0.0	21.1	12.9	21.4	0.1	0.0
Post-basin mm/yr	2.9	357.8	18.0	15.5	313.6	10.3	39.3

Basin salt balance

A basin salt balance shows that about 97% of the salt pumped with the inflow from the drain was lost with the water leaked from the basin. Substituting salt loads into the basin water-balance equation (Equation 6) for the period 22 December 2004 to 31 December 2006 was:

$$23\,600 + 0 = 0 + 600 + 22\,900 + 150 \text{ (t)}$$

The zero values for precipitation and evaporation show that these have no or an insignificant effect on the salt balances compared with the other values. The release of about 32 000 kL of basin storage into the contingency basin in May 2005 took about 600 t of salt with it.

At the end of the monitoring period in December 2006 the basin contained 150 t of salt dissolved in the remaining 3100 kL of water (Equation 6), so the salinity of the remaining water in the basin was 48 973 mg/L (Fig. 40). At this time the average flow-weighted salinity of the drain discharge was 36 000 mg/L.

Basin site salt balance

The basin site salt balance reflects the leakage of 22 900 t of salt pumped into the basin with only 150 t left in storage by the end of the monitoring period. Recirculation accounted for 2800 t of the salt pumped into the basin. Of the remaining 20 800 t transported to the basin site by the drain, 20 600 t leaked into the underlying aquifer. The salt balance was simulated for the entire monitoring period by using the salt loads for the various components in the site's groundwater balance equation (Equation 8):

$$300 + 22\,900 + 0 = 1300 + 23\,000 + 700 + 2800 \text{ (t)}$$

Like the drainage site's salt balance, the basin site's salt balance could not be solved accurately as the mass of salts washed away in runoff after being deposited on the soil surface by capillary discharge was unknown. Of the 23 000 t of salt deposited on the soil surface by capillary discharge, an unknown proportion was lost in surface runoff and/or leached back into the soil.

A uniform groundwater salinity of 37 314 mg/L was used to calculate the salt load for aquifer discharge into the basin site, corresponding with aquifer discharge salinity into the drainage site. The assigned groundwater salinity used to calculate loads for aquifer storage change, groundwater evaporation and aquifer discharge out was 59 536 mg/L. This salinity was equivalent to the average salinity of the recirculation water intercepted by the cut-off drain (Appendix B) and represented the natural and leakage groundwater beneath the basin.

Potential post-basin salt deposition onto the soil surface of the 61 ha basin site was nearly 15 times pre-basin levels (Table 7). The high salt deposition rate reflected that watertables beneath the site were rising and nearer the surface mainly due to basin leakage.

Table 7 Representative annualised basin site salt balances expressed in t/ha/yr

	$A_{D\ IN}$	L	U	ΔA_S	G_E	$A_{D\ OUT}$	G_O
Total t/ha/yr	1.9	159.2	0	8.9	151.9	4.9	18.7
Pre-basin t/ha/yr	4.9	0	0	7.7	12.7	0	0
Post-basin t/ha/yr	1.1	188.9	0	9.2	186.7	6.1	23.4

The 186.7 t/ha/yr of salt potentially deposited on the soil surface by evaporation closely coincides with an equivalent 188.9 t/ha/yr lost by leakage from the basin (Table 7). Although these figures may be coincidence (and the correct deposition rates are not known), increasing masses of salt have accumulated as a white crust on the surface surrounding the basin. During the 2005–06 summer the salt crust is noticeable on the surface in the contingency basin to the south of the main basin, outside the main basin on its other three sides, and mostly within the main basin (Fig. 49).

Figure 49 Basin site from the air showing surrounding surface salt accumulation

6 Discussion

The objective of any drainage scheme for salinity control within the valley floor environment of the Wheatbelt of Western Australia must be to maintain the watertable at a depth where the upward flux of salts by capillary rise is less than the downward flux by leaching or loss with overland flow. To achieve this in an already saline environment the watertable must be lowered or its upward fluxes controlled so that the watertable does not rise to near the soil surface. Of secondary or longer term benefit, a successful drain should provide a pathway for displacing saline groundwater with fresh rain-fed recharge, thereby reducing the salinity of the shallow groundwater.

The potential for benefits from both watertable control and enhanced salt leaching were explored at Morawa by both direct analysis of the watertable and changes in the water and salt balances of the drainage site. Although very preliminary, results suggest that even the small reductions in the watertable height due to the drain could translate to reductions in topsoil salinity. The drain has also been shown to export about 10 000 t/yr of salt from its surrounding aquifer.

Of most importance is the conclusion that unbounded drains (Appendix A) are a less than effective tool for managing shallow watertables to aid in the recovery or protection of saline land. Within the context of its surrounding large aquifer the single channel of the Morawa drain appears to be incapable of sufficiently enhancing the groundwater discharge from the catchment and so lower and control the watertable. This inadequate performance does not appear to be a function of soils unsuitable for drainage but a matter of scale, as discussed below.

This report highlights some of the limitations challenging expectations of retaining saline drain discharge within evaporation and detention basins. Although the discharge water from the Morawa drain was largely contained and evaporated from a basin and surrounding land, the salt remains and will eventually be lost, carried downstream in surface runoff. More work is needed to better understand the broader implications of this compared to the ‘do nothing’ and other discharge management options.

6.1 Drain discharge

Water discharge

Sustained drain discharge associated with virtually no reduction in watertable height is a common result to unbounded drains found elsewhere in the Wheatbelt of Western Australia (Cox & Tetlow in press). An unbounded drain is a single drain with no groundwater boundary (Appendix A) to constrain the movement of groundwater into and out of its drainage site (Section 5.1). In comparison, drains in bounded schemes are constructed parallel to each other so that each functions as a groundwater boundary to those alongside.

When groundwater recharge falls to zero, discharge from bounded drains reduces corresponding with the watertable reduction. During a prolonged absence of rain-fed recharge, the watertable could fall close to the channel floor level and groundwater inflow into the drain could stop. Because the Morawa drain is unbounded, its discharge rate was regulated by the supply of both local rain-fed recharge and aquifer discharge (Appendices E & F). Without rain-fed recharge, inflow to the drainage site and so to the drain were sustained by aquifer discharge from the greater surrounding aquifer.

In Figure 50 the monthly water balance volumes for groundwater inflow (Appendix B), aquifer storage change (Appendix E) and groundwater supply are expressed in millimetres depth equivalent across the drainage site. Groundwater supply is the combination of rain-fed recharge and aquifer discharge that together represent the supply side of the water balance equation (Equation 2). Together, recharge and aquifer discharge represent the total volume of groundwater supplied to the drainage site; that is, the maximum volume of groundwater available to be drained.

If the groundwater inflow rate to the drain exceeded supply to the drainage site, the inflow had to come from a reduction in aquifer storage. Groundwater inflow to the drain which was about 6 mm for January 2005 was sourced from a combination of both aquifer storage reduction (–5 mm) and groundwater supply (2 mm) (Fig. 50). Because rain-fed recharge was 0 mm in January 2005 (Fig. 64) all of the 6 mm of groundwater inflow to the drain was sourced from aquifer discharge and storage reduction with resultant watertable level declines.

The groundwater supply to the drainage site rarely fell below the groundwater inflow rate to the drain (Fig. 50). Most of the time there was an over-supply of groundwater to the drainage site compared with that removed by the drain. This has implications for watertable management (Section 6.2) because if groundwater inflow to the drain is exceeded by groundwater supply, additional water must raise the watertable and/or be lost by capillary discharge.

From drain construction in November 2004 to May 2005, groundwater inflow to the drain exceeded the groundwater supply to the drainage site (Fig. 50). During this period, groundwater inflow to the drain was sustained mainly by groundwater released from aquifer storage (negative values) as watertables beneath the drainage site fell. From May 2005, this situation was sometimes reversed as aquifer storage increased when groundwater supply to the site exceeded the inflow to the drain. Hence, when groundwater supply is equal to or less than inflow to the drain, aquifer storage decreases, and aquifer storage increases when supply is greater than inflow.

The supply to discharge results reflect those embedded in the water balance that show that the drain removed the equivalent of 94 mm depth of groundwater from the drainage site (Section 5.2). During this time the combined supply to the site and aquifer discharge was 114 mm, with the remaining groundwater lost by evaporation.

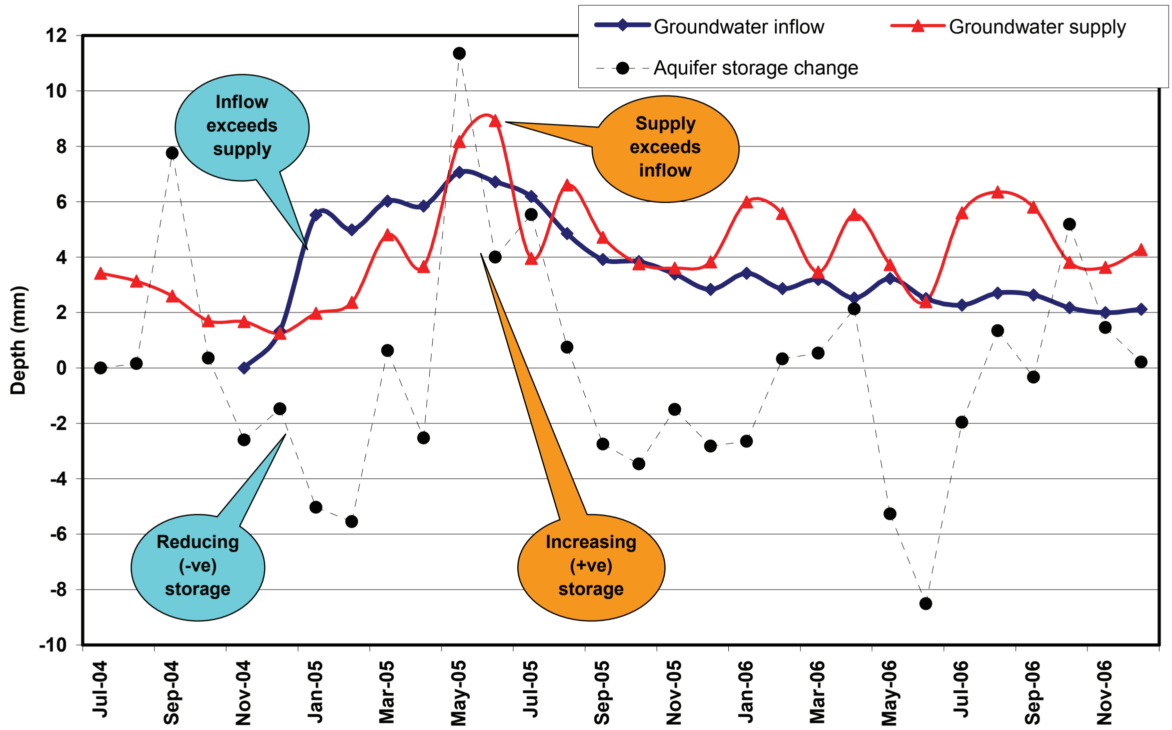

Figure 50 Groundwater supply and storage change beneath the drainage site and groundwater inflow to the drain

Paradoxically, when the drain was most overloaded by groundwater supply, it was most efficient at removing water but most inefficient at controlling the watertable height. Because increased drainage efficiency is related to bigger head difference between the watertable and drain floor, watertables must rise to increase the groundwater inflow rate to the drain (Appendix C).

In 2005, 62 mm of groundwater drained from the drainage site; this was equivalent to the amount supplied to the site, whereas in 2006 only 34 mm of the 60 mm supplied drained. The main differences between these two years were the intensity and timing of the recharge and subsequent watertable elevations. In 2005, over 50% (12 mm) of the annual rain-fed recharge (Appendix E) occurred in May and June causing the watertable to rise by an average 0.5 m (Fig. 51). The head rise substantially increased the groundwater inflow to the drain, enabling it to remove a larger proportion of the groundwater. The higher watertable also occurred in winter when the potential for groundwater loss by capillary discharge and evaporation was at its lowest (Fig. 2). The drain had more time to remove the groundwater that might otherwise have been lost by capillary discharge and evaporation.

The 2006 groundwater supply of 60 mm was more evenly distributed through the year (Fig. 50) so did not raise the watertable head and subsequent inflow to the drain. Accordingly, a lower proportion of the groundwater drained compared to 2005. These two post-drain performance scenarios are represented in Figure 51 where the cumulative groundwater inflow falls below the cumulative groundwater supply in early 2006. From November 2004 to January 2006, cumulative groundwater inflow to the drain exceeded cumulative groundwater supply to the site because the higher watertable head was sufficient to drive the higher rate of groundwater inflow into the drain. Overall, groundwater supply was less than groundwater inflow because a portion of groundwater supply was also derived from reducing aquifer storage during this period.

The groundwater inflow rate began to fall in response to reducing groundwater heads during late 2005 (Fig. 51). The increasing gap between cumulative groundwater inflow and supply represents the increasing proportion of the groundwater beneath the drainage site being lost by capillary discharge during 2006.

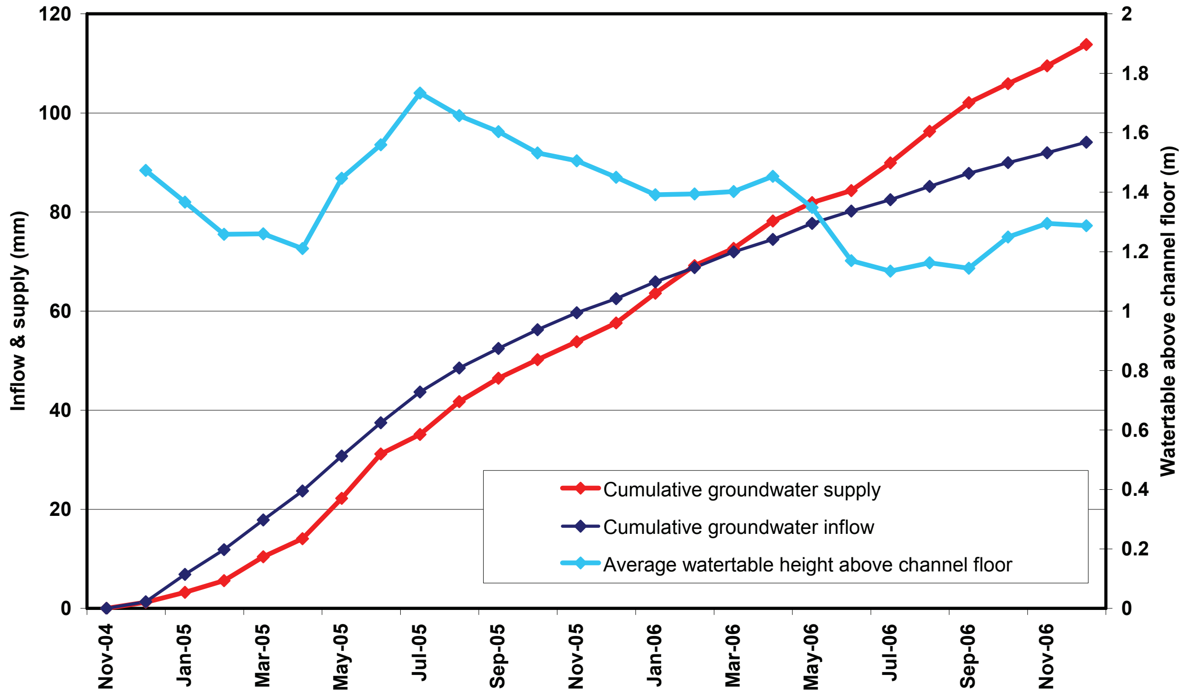

Figure 51 The relationship between groundwater inflow and supply with changing watertable head

The resultant discharge rate of the drain at any given time could be reduced to the outcome of the relationships between the timing and intensity of recharge (supply) events in relation to the height of the resultant watertable. The results of this relationship then determine the proportion of water to be drained to the proportion that will continue to be lost by capillary discharge.

Because the timing and intensity of a physical parameter responding to a notional one is captured by the AMRR trend that partially 'drives' groundwater supply and head, drain-flow was 'responsive' to AMRR (Fig. 33). The drain-flow appeared more responsive to the AMRR trend when the head difference between the watertable and drain floor was greater. As the head difference decreased, the drain became less responsive to AMRR trends until the resultant rain-fed recharge significant raised the channel floor–watertable head.

In summary, the Morawa drain provides a groundwater outlet for the catchment where before there was none. The rate at which the drain removes rain-fed recharge and aquifer discharge depends on the surrounding groundwater heads at the time, with the whole groundwater system, including the drain, responsive to AMRR trends. The drain and capillary discharge both remove groundwater, with the proportions removed dependent upon previous recharge and current groundwater levels.

Salt discharge

Simple groundwater salt mass balance calculations show that enormous masses of salt have accumulated beneath the drainage site. The groundwater salinities and aquifer thicknesses of the deep bores in transects 1–3 provide for an estimated salt storage of 939 000 t (1574 t/ha) within the aquifer beneath the drainage site. This estimate was based on an average groundwater salinity of 26 000 mg/L and soil groundwater storage of 30%.

Regardless of drainage, aquifer discharge transports dissolved salts from beneath the surrounding elevated landscape towards the valley floor and into the drainage site. Prior to drainage the aquifer discharge salt load was estimated to be 6268 t/yr (10.5 t/ha/yr) and post-drainage 8000 t/yr (13.4 t/ha/yr) (Table 4). The post-drain increase in aquifer discharge was just as likely caused by natural variability in groundwater levels surrounding the drainage site as anything to do with drain installation.

Within the short-term context of the salt balance equation for the drainage site (Section 5.2), the mass of salt derived from marine aerosols as salt fall (Section 4.4) and transported in surface water run-on was considered as negligible. Although it must be pointed out that run-on and runoff are responsible for the dissolution and mobilisation of many tonnes of salt both to and from the soil surface of drainage site, these salt loads have not been quantified at Morawa.

Under pre-drain conditions, surface water runoff was considered as the only process exporting significant loads of the dissolved salts that had been transported to the soil surface by capillary discharge. Of the estimated 15.9 t/ha/yr of salts transported to the surface by capillary discharge (Table 4), what proportion was exported from the drainage site in surface runoff in any given year is unknown. The export rate would probably be vary widely, associated with the timing and magnitude of the infrequent runoff events from the drainage site and surrounding valley floor. Some of the watertable responses show salinity trends that reflect the re-leaching rather than the export of accumulated surface salts, in response to rainfall (Section 4.2).

Attempts to estimate post-drain salt export in surface runoff from the drainage site faced the same limitations as the pre-drain estimations. Although post-drain capillary discharge of salts fell to an average of 5.5 t/ha/yr, it is unknown how much of this was exported in surface runoff from the drainage site during the monitoring period.

The only known value for salt export from the drainage site was the post-drain salt load of 10 388 t/yr (17.4 t/ha/yr) discharged by the drain (Table 4). Being sourced from groundwater inflow to the leveed drain, it is known that this salt-flow originated from the aquifer and was exported from the drainage site by the drain.

At the drain-flow export rate of 10 400 t/ha/yr, all the salt could theoretically be exported from beneath the drainage site in just over 90 years; that is, with the drain removing an equivalent of 17.4 t/ha/yr from the 1574 t/ha accumulated beneath the drainage site. As aquifer discharge delivers an average 13.4 t/ha/yr into the drainage site, net salt export reduces to 4 t/ha/yr and the theoretical salt leaching time for the drainage site rises 393 years.

The reality is that aquifer discharge reflects the connectivity between the drainage site and surrounding 1080 km^2 of catchment (Section 2.3). Because aquifer discharge will continue to deliver salts into the drainage site, any decline in the salinity of the aquifer underlying the drainage site can only be to a level commensurate with the salinity of aquifer discharge. Unless parallel drains are installed to achieve a greater disconnection between the shallow groundwater system that can be influenced by the drain and deeper groundwater aquifer discharge the potential to achieve beneficial shallow groundwater salinity reductions is limited.

6.2 Groundwater change

Watertable height

The Morawa drain is now part of the catchment valley floor groundwater system and in conjunction with capillary discharge maintains watertables below the ground surface. Without evaporation by capillary discharge and/or drainage, groundwater would rise to the surface and generate seepage across the valley floor.

Before drainage, there were gentle hydraulic gradients on the watertable towards the centre of the valley floor and where the drain is now constructed (Section 4.1). These gradients largely mirrored those of the land surfaces from which groundwater evaporated. The uniformity of the vertical movements of the watertables indicate that:

1. recharge to, and capillary discharge from, the watertable was relatively uniform and/or
2. the underlying valley floor aquifer is conductive and relatively unconfined, allowing groundwater to migrate freely to redistribute localised head differences in the watertable, i.e. remain flat.

Although the first statement is considered true, the watertable, drain and basin responses measured and discussed in this report more closely reflect the aquifer conditions of the second statement. Under these conditions, a significant fall in watertable solely in response to the unbounded drain at Morawa is unlikely, largely because groundwater movement towards the drain is thought to be radial (Appendix F) through the underlying aquifer, allowing groundwater to migrate freely towards the drain or away from the basin.

This response to the unbounded drain coupled with the aquifer conditions that promote radial flow and offer little resistance to the redistribution of groundwater can be used to explain the shape and extent of the drawdown curves alongside the Morawa drain. In response to drainage, watertable drawdowns extend to large distances horizontally but only small distances vertically. Only where the radial groundwater flows converge near the channel floor does the resistance to groundwater flow increase with head lost, and the watertable fall (Appendix F; Fig. 65). Unfortunately, at Morawa, the convergence and associated significant vertical watertable drawdown only occurs within tens of metres of the channel (Figs 14 & 16).

The effect of the drain on watertable height beyond the tens of metres from the channel is minor and remains inconclusive. The relationships between the drain, groundwater heads, groundwater supply and potential for capillary discharge all contribute to how much groundwater will be drained and lost by capillary discharge, and the resultant watertable levels. If the pre-drain valley floor groundwater levels were controlled by an equilibrium between the inputs of recharge/aquifer discharge against capillary discharge from the watertable (Section 5.1), any post-drain effect on groundwater levels could go undetected.

Without a water balance, the drain effect on the watertable could go undetected because in installing the drain the main process of groundwater level control gradually changes from capillary discharge to drainage. This shift was detected because there were pre- and post-drainage estimates of groundwater loss by capillary discharge.

In effect, the water balance (Equation 2) suggests that installing the drain resulted in a small reduction in average watertable height with an associated reduction in capillary discharge from a rate equivalent to 43 to 15 mm/yr. The inference is that the 28 mm/yr difference between the pre- and post-drain results represents water removed from the drainage site by the drain.

To achieve this shift from capillary discharge to drainage, the watertable needed to fall by an estimated average of 0.29 m across the site. This value was calculated by using the ‘watertable depth to rate of capillary rise’ chart (Fig. 67) in reverse to calculate watertable depth from the average annual capillary rises. To achieve the pre-drain capillary discharge of 43 mm/yr, average watertable depth needed to be 1.07 m. The pre-drain average monthly measured watertable depth was 1.01 m. From the chart, the post-drain capillary discharge was 15 mm/yr with 1.36 m watertable depth although the measured average post-drain watertable was 0.24 m higher, 1.12 m bgl.

Bearing in mind that the water level results from the chart (Fig. 67) are capillary rise and not capillary discharge, the actual water levels would need to be higher than those results to achieve the water balance capillary discharge rates.

The changing proportions of capillary discharge to drainage illustrate the sensitivity of capillary rise to even minor level changes once the watertable is within about 1 m of the surface. When the capillary fringe reaches the soil surface, small changes in the underlying groundwater level can have profound effects on soil water content and so increased potential for recharge (Gillham 1984). Recharge and capillary rise cause associated changes in the potential for topsoil waterlogging, capillary discharge, and, where groundwater is saline, topsoil salinity.

The subtlety of these topsoil responses to watertable changes underpins anecdotes that previously saline areas of land can be cropped again during dry years in response to even natural watertable decline. The small watertable reductions in response to the Morawa drain also conform with anecdotes suggesting that drainage can improve productivity of once saline land without large falls in groundwater level. However, in both of these cases, the risk is that with wetter seasons, the productivity of the land is not assured.

Watertable and topsoil salinity

There was no evidence at Morawa of the expected declining watertable salinity in response to the enhanced capacity for leaching and the displacement of saline groundwater provided by the drain (Section 4.2). As discussed above, because aquifer discharge and evaporation remain significant post-drain components of the water balance, the capacity for short-term reductions in watertable salinity is limited. The ongoing contributions of salt by aquifer discharge, the concentration effects of evaporation plus the existing high salt storage levels beneath the valley floor will combine to overwhelm any groundwater dilution effect of ‘fresh’ recharge well into the future.

Topsoil salinity could reduce within the short time span of this project with salinity reduction responses. Capillary discharge is seen as the main process by which salts are transported from the watertable to the topsoil and soil surface (Cox & Tetlow in press). Any capillary discharge reduction would lower the surface salt accumulation rate. The groundwater salinity also affects the deposition rate. Higher salinity groundwater contributes more salt per unit volume of capillary discharge and so results in higher deposition rates than fresher groundwater.

Post-drain potential accumulation rates were lower than pre-drain rates. Calculated capillary discharge (Appendix G) and groundwater salinity measurements (Section 4.2) produced an estimated annual pre-drain salt deposition of 7250 t. The estimated annual post-drain salt deposition within the drainage site was 5350 t. The 2006 post-drain deposition was about half of the 2005 value, in response to the lower average 2006 watertable levels (Fig. 52).

Calculated monthly deposition rates closely corresponded with changes in watertable depth, reaching a maximum of about 1900 kg/ha/month, when the average watertable depth was about 0.8 m beneath the drainage site (Fig. 52). The approximately 0.5 m greater average watertable depth (1.3 m) for the comparison bores (Section 4.1) meant the watertable levels fluctuated less in response to recharge, and the capillary discharge rate and deposition rate varied less than the drainage site rate. As a consequence, the corresponding September 2004 maximum salt deposition was 1500 kg/ha/month surrounding the comparison bores (Fig. 52).

The average monthly post-drain salt deposition rate was 74% of the pre-drain rate. From July–December 2004 average pre-drain deposition was 1012 kg/ha/month with a range 653–1218 kg/ha/month (Appendix CD 6.2). The average post-drain 2005–06 salt deposition was 746 kg/ha/month, with a range 315–1882 kg/ha/month. The lower post-drain rates, during early 2005 and late 2006, were associated with concurrent lower watertable levels (Fig. 52).

The 2005–06 average monthly salt deposition rates surrounding the comparison bores were 3.2 times higher than for the preceding July–December 2004. These increased rates that coincided with the post-drain 2005–06 period were associated with the elevated and then slowly receding groundwater levels in these bores at the time (Fig. 12). After the significant May–June 2005 increase in their groundwater levels, subsequent water levels did not fall to or below their December 2004 levels until about August 2006. In comparison, most of the groundwater levels beneath the drainage site fell back to or below their December 2004 levels before the end of 2005 (Section 4.1).

By using the capillary discharge rates calculated from Equation 2 rather than Equation 1 (Section 5.2) the post-drain 2005–06 average annual salt deposition rate is reduced from 5350 to 879 t. At an equivalent average rate of 132 kg/ha/month, post-drain salt deposition was about 26% of the 464 kg/ha/month pre-drain rate.

The alternative approach using Equation 2 results resulted in the most significant reduction in the potential salt deposition rate during the May–October 2005 period of high watertable compared to the calculated capillary discharge results (Fig. 52). Even though the post-drain watertable rose 0.1 m higher than the pre-drain September–November 2004 level, potential post-drain salt accumulation peaked at 304 kg/ha/month compared to the pre-drain 719 kg/ha/month.

The difference between the peak pre- and post-drain potential salt accumulation rates could reflect the real difference in the potential capillary discharges at the time. Potential evaporative loss, a driver of capillary discharge, was on average 3.8 mm/d for May–October 2005, about 50% of the September–November 2004 pre-drain period. These differences in potential rates highlight the significance of the timing of as well as the magnitude of fluctuations in the watertable nearer the soil surface in contributing to the potential for soil salinity.

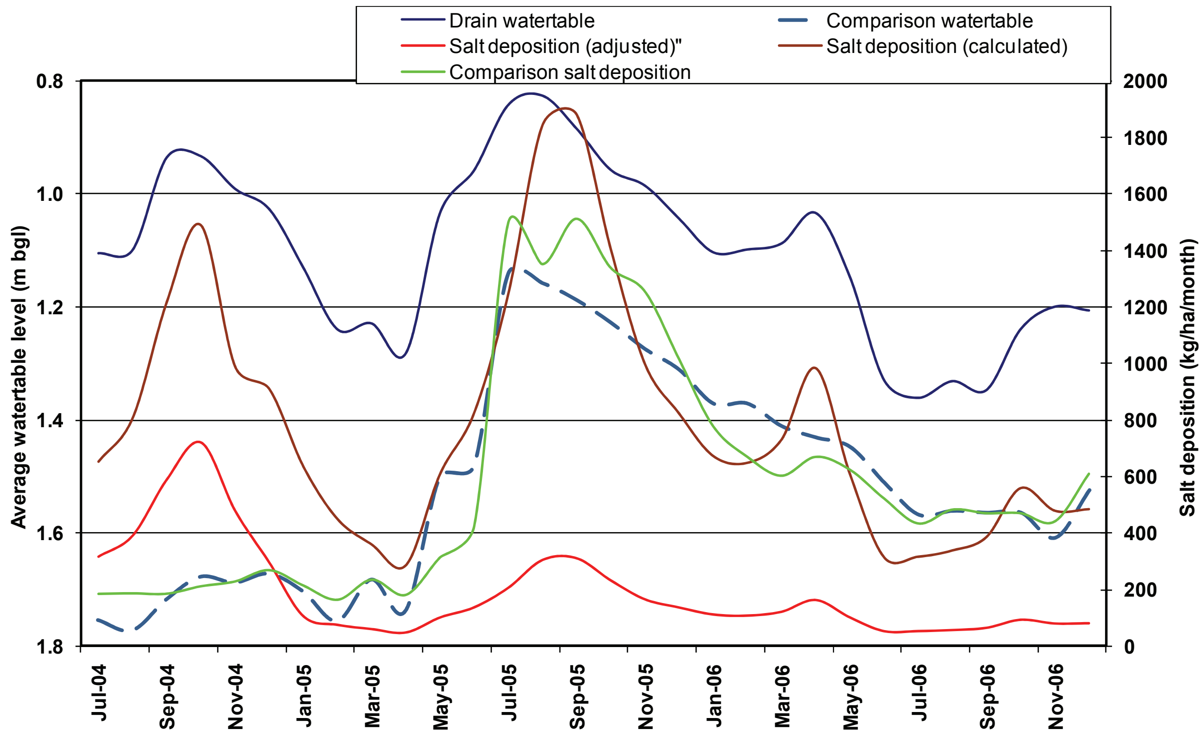

Figure 52 Average monthly watertable levels and capillary salt discharges for the drainage and comparison sites

Although only part of the story, the potential salt deposition rates discussed above suggest that the drain's contribution to lowered watertable levels could have a substantial impact on topsoil salinities. Mean soil salinities from 22 samples taken along the transect bore alignments show a 50 mS/m (275 mg/L approx.) reduction in topsoil salinity of the upper 15 cm of the profile between January 2005 and October 2006 (Bell et al. 2009). The 2006 topsoil salinities were about 200 mS/m (1100 mg/L approx.) lower than the average of the samples taken adjacent to three of the comparison bores in October 2006. There were no 2005 samples collected from the comparison sites.

Although there is potential for *seasonally* positive outcomes from the drain in terms of reducing topsoil salinity, the inability of the drain to demonstrate global control of the watertable means these benefits may be localised and short lived. Even if the drainage was effective at lowering the watertable, using capillary discharge and salt accumulation rates to assess potential benefits remains an area requiring considerable development.

6.3 Making drains work

The drain and aquifer characteristics at Morawa have demonstrated that substantial volumes of groundwater can be drained from the landscape. So, the configuration of the single unbounded drain layout is thought to be the reason that the drain is unable to lower and control the watertable, not the aquifer and soil conditions. This raises the question: If multiple (parallel) bounded drains are needed, how far apart could they be while providing adequate control of the watertable between them?

Computer modelling using the Hooghoudt Steady State Equation (Ritzema 1994) and the drain variables identified and calculated in this report show that 2.5 m deep bounded (parallel) drains can adequately control the watertable at 1.5 m below ground level. To control the watertable to this depth, the spacing between the drains depends on the volume of recharge to be drained. The greater the recharge, the closer the required drain spacing.

The depth of recharge is the design recharge, otherwise referred to as the drainage coefficient (Ritzema 1994), which is the expression of the depth of recharge to be drained within a given period of time. Although designing for a lower drainage coefficient will result in wider drain spacing, this can be risky. If real recharge exceeds the drainage coefficient, the watertable between the drains will rise above the design maximum watertable level; in this case, 1.5 m below ground. Parallel drain design requires designing the optimum drain spacing to manage the drainage coefficient balanced against the risks associated with the timing and frequency of recharge events that could exceed it and cause crop losses.

Using the results from Morawa, drain spacing design could be based on three recharge scenarios that reflect different levels of risk:

- If there was a high risk of crop loss associated with groundwater rise due to recharge, the drain spacing design could be based on the highest monthly drainage coefficient. In June 2005, combined aquifer discharge and rain-fed recharge produced a drainage coefficient of 9 mm/month (Appendix CD 6.1), expressed as an annual equivalent of 108 mm/yr. At a steady-state drainage coefficient of 108 mm/yr a drain spacing of 270 m would maintain the watertable at a minimum of 1.5 m below ground.

- A drain spacing designed using a drainage coefficient of 60 mm/yr would provide a moderate amount of protection from groundwater rise. This drainage coefficient is equivalent to the post-drain groundwater supply to the drainage site (Section 6.1) and the drain spacing could be increased to about 370 m.

- The drain spacing could be further increased to 660 m if the drain were designed for a drainage coefficient of 21 mm/yr, equivalent to the average post-drain rain-fed recharge to the site (Section 5.2). Watertable control at this spacing could only be achieved in the absence of aquifer discharge. The watertable between the drains would rise above 1.5 m below ground level in response to the May–June 2005 measured recharge. Further modelling and analysis could be undertaken to estimate the frequency of such recharge events and the associated risk of crop losses.

At 660 m parallel drain spacing, a parallel drain is only expected to provide marginal watertable control that could lead to the recovery of saline land for barley cropping in most years. Even at this spacing, the watertable zone of influence of each drain is 330 m, 60 m less than the interpreted watertable zone of influence; the distance used to define the perimeter of the drainage site. These results reflect that, in comparison to the three scenarios above, there is very little potential for the existing drain to provide adequate watertable control at or below 1.5 m below ground level within the drainage site.

The need to maintain the watertable lower than 1.5 m below ground level is to enable barley cropping on the recovered saline land. For wheat cropping, the saline watertable needs to be controlled to a greater depth, at least 1.8 m deep (Nulsen 1981). Wheat cropping would require even closer drain spacing than barley cropping.

6.4 Basin performance

Leakage

Evaporation basins are used around the world to divert saline irrigation drainage water from rivers, lakes and other natural water bodies to protect downstream water users and natural ecosystems. The techniques of constructing evaporation basins in Western Australia are similar in concept to those in other states, but not in scale and application.

Evaporation basins are broadly categorised as 'sealed' or 'leaky' (George & Nott 2004). Sealed basins depend on the low permeability of the basin floor to prevent leakage that might pollute underlying groundwater systems. These basins may be lined with impermeable clays or plastic liners to prevent leakage. Over time the concentrated salts need to be removed from the basin to maintain its evaporative efficiency. The leakage rate from leaky basins is controlled by the characteristics of their underlying aquifers. Leakage rates of 0.5–1 mm/d are acceptable and provide the beneficial leaching of salt from the basin waters to the underlying aquifer. The continual leaching of salts maintains the evaporative efficiency of the basin by preventing concentration of salts.

Basin leakage rates greater than 3 mm/d are undesirable as the interception and recovery of leakage may become unmanageable (George & Nott 2004). The Morawa basin was designed as a leaky basin, but it produced unmanageable leakage rates up to 30 mm/d. Groundwater levels alongside the cut-off drain rose soon after basin filling started in December 2004 (Fig. 21). The unmanageably high leakage rates into the cut-off drain, intended to intercept the leakage, led to the cut-off being disconnected from the main drain in June 2005.

As at Morawa, the watertables in most of the valleys in the eastern Wheatbelt are within 2 m of the surface. The low gradient along these valleys (approximately 1 m/km) results in a slow lateral groundwater flow of 1–10 m/yr under natural conditions. Shallow watertables and slow lateral movements hinder leakage from a basin; the leakage is controlled only by the hydraulic characteristics of the aquifer under the basin and the resulting increased hydraulic gradients (Leaney & Christen 2000).

The inability of the basin leakage to substantially increase the hydraulic gradients of the surrounding watertable confined the spread of the leakage to largely within the 61 ha basin site at Morawa (Section 5.1). The pre-basin watertable of the basin site was about 1 m below ground level, and groundwater seeped into the 0.75 m deep burrow areas during construction. With less than 1 m between the pre-basin watertable and soil surface, the hydraulic gradient was vertically confined, reducing the rate of groundwater movement away from the basin. As the watertable rose in response to leakage, increased groundwater evaporation and capillary discharge around the basin site progressively removed groundwater (Section 5.3) as it moved away from the basin. This progressive loss is partially responsible for hydraulic gradients sloping away from the basin (Figs 20 & 22).

Nearly 80% of the water pumped into the Morawa basin leaked via the underlying aquifer to the surrounding site (Section 5.3). The leakage was exacerbated by the water held above the natural ground level. More precisely, the leakage was exacerbated by increasing the head difference between the basin water level and its surrounding watertable. The infiltration capacity of the 5 ha basin floor and the transmissivity of the aquifer beneath allowed for an average leakage rate of 11.7 mm/d. The peak leakage of 30 mm/d occurred during June 2005, while the basin was filled to its maximum storage volume and depth (Fig. 47).

The average leakage rates for 2005 were 16.2 mm/d and 7.0 mm for 2006. The difference in these and any other variations in leakage can be largely explained using the relationship between the water level head in the basin and the transmissivity of its underlying aquifer, with the Darcy equation:

$$Q = K\ (\Delta H/\Delta V)\ A \qquad \textit{Equation 9}$$

Where: Q : The volume of leakage per day

K : Hydraulic conductivity of the aquifer of 1.2 m/d (Section 2.5)

($\Delta H/\Delta V$): Average hydraulic gradient between the basin water level and groundwater level at 20 m NW in Transect 4 bore 019

A : Cross-section of the aquifer which in this case is represented by the 950 m basin wall multiplied by a 20 m aquifer thickness, from bore 19d (Appendix CD 2.5a)

The 950 m basin wall length was used to represent the width of the aquifer cross-section because it is the most restricted section of the aquifer through which all leakage had to pass when moving from the inside to outside of the basin perimeter. When multiplied by the 20 m aquifer thickness, the cross-sectional area of the aquifer was 19 000 m^2.

With an average head difference of 0.53 m between the basin water level and groundwater level in bore 019, average daily leakage was calculated as:

$$604 = 1.2 \times (0.53 / 20) \times 19\ 000$$

An average leakage of 604 kL/d over two years is 441 000 kL, which is close enough for planning and design purposes to the basin water balance leakage rate of 433 000 kL (Section 5.3). The average 0.85 m depth of water in the basin produced an average head difference of 0.7 m between the water levels in the basin and bore 019 during 2005. At 0.7 m of head, the calculated leakage rate for 2005 was about 291 000 kL, which compares favourably to the water balance result of 296 000 kL (Appendix CD 5.9). The calculated leakage of 291 000 kL from the 5 ha basin area is equivalent to an average depth of 15.9 mm/d compared to the 16.2 mm/d average water balance leakage rate for 2005.

The calculated leakage for 2006 was about 1.2 times the water balance result, at 149 000 kL. This was possibly caused by the less frequent groundwater measurements and subsequent inaccurate average head, or reduced leakage reflected in the water balance by the basin sometimes drying out. Reducing the groundwater head from the measured average of 0.36 m to 0.31 m produced the same leakage result as for the 2006 basin water balance. The peak leakage of 30 mm/d, in June 2005, occurred with 1.2 m depth of water in the basin and a head difference of 1.18 m between the basin and bore 019 water levels.

To maintain leakage rates at or below the acceptable level of 1 mm/d requires that the average head difference between the basin water level and surrounding watertable does not exceed 0.045 m. Based on the average depth of water evaporated from the existing basin, and allowing for 1 mm/d of leakage, a 20 ha basin would be needed to evaporate the discharge from the Morawa drain.

Increasing the basin size from 5 to 20 ha results in increasing the wall length with an associated increase in aquifer cross-sectional area, aquifer transmissivity and so increased potential for leakage. To reduce this leakage by design and reduce transmissivity and leakage rate, the basin would need to be enlarged further to reduce the head difference between the waters inside and outside. Very quickly, the required basin area increases to over 30 ha, close to the combined 31.6 ha area of the project's primary and contingency basins.

Even by distributing the water over a 30 ha land surface area the capacity to control the leakage from an enlarged basin to prevent surrounding watertable rise was limited by the aquifer transmissivity at Morawa. To enclose a 30 ha basin requires about 2160 m of wall. If the watertable were to be maintained at 0.70 m below ground level, outside of the basin, about 1950 kL/d of leakage would potentially leak from the basin under the wall and would need to be recaptured.

When the 32 000 kL of water was released from the Morawa primary basin into the contingency basin, in July 2005, the water rapidly infiltrated the contingency basin floor and raised the watertable inside and outside the contingency basin. The area affected by this higher watertable is clearly evident in Figure 49 by the (white) salt deposited on the soil surface to the south of the primary evaporation basin. Given the extent of the land area and leakage rates, ponding water above the surface for evaporation could not be sustained within the contingency basin.

The basin responses at Morawa demonstrate that in transmissive aquifer conditions basins need to be sealed or excavated to below the watertable if leakage rates are to be managed. The suggestion is that basin configurations need to be more like the region's saline groundwater lakes. To achieve this at Morawa would require excavating about 270 000 m^3 of earth for the 30 ha basin area and an almost 10-fold increase in construction costs compared with costs of the current basins.

If sufficient land was available, an alternative approach could be to improve upon the experiences from the project by ensuring that leakage is not lost from the basin site. Rather than being close to the basin, a cut-off drain could be placed at the outer extent of any likely groundwater rise in response to basin leakage. At Morawa, the cut-off drain could be placed around some or all of the perimeter of the basin site to prevent the further expansion of the groundwater mound. The cut-off drain need only extend to just below the pre-basin groundwater levels, allowing leakage and the leaching of salt from the site to continue at natural rates.

Evaporation

The 101 800 kL (Equation 6) of evaporation from the 5 ha basin surface area was equivalent to 38% of the pan evaporation measured at Station 508042 (Section 2.1) for 2005–06. This figure was lower than expected as at various times throughout the project the basin was not full and the water surface area was smaller than 5 ha (50 000 m^2).

During 2005, the average surface area of the water was about 32 600 m^2 from which evaporated 65 900 kL of water, so free evaporation was therefore 2023 mm, or 78% of pan evaporation. During 2006, 35 800 kL evaporated from an even smaller average water surface area of 14 200 m^2, at a rate equivalent to 91% of the 2781.7 mm pan evaporation.

The inaccurately high percentage of evaporation for 2006 was possibly the result of calculation errors in the water surface areas or that substantial evaporation occurred from the wetted but not inundated soil surfaces within the basin. Both of these are possible because the water surface area was more sensitive to water level changes during 2006 as it fluctuated between the borrow areas and natural ground surfaces in response to small changes in water level (Section 4.7). Being close to the water levels would mean that the natural ground surfaces in the basin were saturated, though not inundated, allowing enhanced evaporation from the soil surface.

Salt

The high leakage rate from the basin largely prevented the evapoconcentration of salts within the basin waters (Section 4.7). There was no evidence of groundwater salinity changes in response to saline water leaking from the basin because the basin waters and the underlying aquifer had similar salinities (Figs 27 & 40).

The leakage of 22 900 t salt (Section 5.3) from the basin was equal to about 14% of the estimated pre-basin salt storage of 176 000 t beneath the basin site. The post-basin salt balance provided no confirmed results on the effects of the additional salt inputs to the basin site, other than to suggest it led to an increased storage of just over 1100 t which was directly related to the salt load of increased aquifer storage beneath the basin site. It does not reflect that salts may have accumulated or were stored within the unsaturated soil profile and on the soil surface.

The salt balance suggested that the majority of salt was or will be exported from the site by capillary discharge of salts to the soil surface, and subsequent removal by rainfall-generated overland flow. For this to occur, salts must pass through the unsaturated soil profile and onto the soil surface. So, salts lost by capillary discharge refer to salts somewhere in transit between the watertable, the upper soil profile and being washed from the site. Neither soil salinity nor discharge and salt load measurements from the site were conducted to confirm the existing situation.

An estimated 14-fold increase in capillary discharge-driven salt export was needed after basin construction to maintain the salt balance between the salt inputs to and outputs from the basin site. The calculated (Equation 7) potential salt export increased from 10.6 t/ha/yr to 88.9 t/ha/yr in response to the increased potential for capillary discharge caused by the rising post-basin watertables. At an average export rate of about 5400 t/yr, the 88.8 t/ha/yr was as yet insufficient to provide for the 14-fold increase in the salt load that needed to be exported from the site.

The calculated salt load from the restricted basin site water balance (Equation 8) was estimated at 12.7 t/ha/yr prior to the basin, increasing to an average 186.7 t/ha/yr after basin construction (Table 7). To provide for the post-basin increased capillary discharge-driven salt load, capillary discharge was 314 mm/yr (Table 6). At this increased export rate about 11 400 t/yr of salt was potentially lost from the site. This was sufficient to export the average 11 450 t/yr basin leakage into the basin site (Section 5.3).

As mentioned above, these should be regarded as suggested values only because they were not confirmed by measurements of soil salt storages and export rates. The enhanced groundwater evaporation potentially transported more salt to the land surface surrounding the basin where it could be removed from the catchment in surface runoff (Fig. 49). Ongoing use of the basin is likely to result in rising groundwater salinity beneath it and its surrounding site until there is equilibrium between the salt flow into the basin and loss through leakage, capillary discharge and surface runoff. The preliminary results from the basin salt balance suggested an equilibrium salt export rate of 186.7 t/ha/yr or 11 400 t/yr from the basin site.

6.5 Construction and running costs

Excluding all of the measurement equipment and associated infrastructure, the main cost components to a landholder implementing the Morawa drainage scheme in 2005 would be:

- Site investigation and approvals: $5 000 (est)
- Drain excavation: $55 500
- Culvert supply and installation: $6 000
- Evaporation basin earthworks: $48 000
- Cut-off drain and sump: $7 200
- Diesel lift pump and motor: $15 000
- Diesel for lift pump (2 years): $12 000 (est)
- Sundries and maintenance (est.): $6 000

The $154 700 total cost is exclusive of GST and includes estimated fuel costs of $1.00/L for the lift pump at the end of the drain. The average fuel consumption for the lift pump was 4 L/h. The total cost does not include the section of drain cut using the rock saw in September 2009. This component was experimental in nature, not essential to the overall performance of the scheme and was directly funded by the landowners.

7 Conclusions

The construction of large 'single drains' for lowering watertables to recover a saline Wheatbelt valley floor at Morawa has provided some benefit in combination with reduced rainfall and recharge effects. Although draining a lot of groundwater the single unbounded drains have not demonstrated the capacity to lower and control watertables in a manner that provides a reasonable degree of certainty that cropping conditions will remain suitable. Unbounded drains are unable to provide farmers confidence that drainage expenditure may be recouped from the profitable cropping of recovered saline land.

The movement of groundwater into Wheatbelt drains has traditionally been portrayed as consisting of shallow horizontal flow, with their associated limited watertable control benefits attributed to low soil permeability. Unfortunately, the majority of historical drainage research in the Wheatbelt focused only on watertable responses. Only when poor watertable responses are measured alongside drains producing high discharges is one forced to investigate the validity of this commonly held opinion.

This report demonstrates that the less-than-anticipated watertable responses to the Morawa drains were not caused by impermeable soil conditions or low soil permeability since large volumes of groundwater discharged from the drain. In fact, higher soil permeability would probably resulted in much more water emerging from the end of the drain, given the potential size of its groundwater catchment.

The inability of the Morawa drain to control the watertable was largely due to the high transmissivity of its underlying aquifer and the volume of groundwater supply to the drain from its surrounding catchment. If, in future, drainage design were based on investigations that included assessment of the underlying aquifer and groundwater catchment, single drains would no longer be the recommended approach for lowering and controlling watertables.

Without 'design' of the recharge–discharge relationship between the aquifer and the drain, drains like this one at Morawa can produce large (even unmanageable) volumes of discharge while unable to control the adjacent watertable for what will probably be very little sustainable benefit in terms of land recovered or protected from salinity in the longer term.

The highly transmissive aquifer conditions that are a problem for the single Morawa drain could greatly enhance the performance of parallel bounded drains at this site and increase the potential for radial groundwater flow that, combined with the low recharge rates, could allow for wide spacings of hundreds of metres between parallel drains.

The high degree of uncertainty surrounding the relationship between the groundwater system and unbounded drain discharge challenges the normal techniques used in predicting and managing discharge. The discharge rate from conventional bounded agricultural drainage schemes can be calculated with a high degree of certainty as it is usually most influenced by rain-fed recharge. There is ample design information to assist in linking such schemes to adequately sized evaporation basins for disposal of their discharge (Singh & Christen 2000).

The anticipated Morawa drain discharge volume was relatively unknown before construction. Of great benefit to the project was the large 'sacrificial' area of saline land at the downstream end of the drain that allowed for the construction of works to retain drain discharge on the property. Ultimately, the lack of gradient, not the works undertaken, was largely responsible for containing the discharge water.

The leakage from the Morawa basins highlighted the importance of understanding the aquifer conditions that underlie any 'leaky' detention or evaporation basins. Aquifer transmissivity was again found to be the most significant factor that affected the leakage rate from the Morawa basins. It was demonstrated that, however shallow the basin, if the water levels inside were much higher than the groundwater levels outside, basin leakage could become unmanageable.

Although the options to restrict basin leakage involved plastic liners or excavating the basin to below the watertable, the consequences of the leakage at Morawa did not appear significant. Because the hydraulic gradient was very low and the evaporation rate high, the spread of the leakage was limited to an area of about 60 ha surrounding the basin. It was considered that placing shallow drains around the perimeter of this area would ensure that any further spread of leakage was contained.

Very preliminary estimates showed the potential for a short-term increase in the salt export rate from the combined basin and drain sites in response to the project. The drain affected an area of about 600 ha, causing approximately 6000 t/yr (10 t/ha/yr) reduction in salt load from the surface of the drainage site. In response to groundwater rise, the potential corresponding salt load from the 60 ha basin site increased by 10 500 t/yr (175 t/ha/yr) with the difference being a net increase of about 4500 t/yr in salt export from the project site.

The complexity of the interactions between the drain and surrounding groundwater systems made it extremely difficult to separate drainage effects from climatic impacts on the watertable, within both the drain and basin sites. The conclusion was that, because the watertable level was affected but not controlled by the drain, the drain will add to the impacts of climate to deliver faster and greater watertable decline than would occur without the drain.

Combined with a drier-than-average 2006 season, the drainage seemed to enable some once saline land to again produce cereal crops at Morawa (Fig. 53), showing that, with a lower watertable, some previously saline land could become productive again. The challenge for drainage in this environment is to retain the productivity of the landscape regardless of seasonal variability.

Figure 53 Previously saline lands adjacent to the Morawa drain

Appendix A Bounded and unbounded drains

The Morawa drainage scheme consists mostly of single channels constructed at or close to the lowest alignments along the valley floors. As such the drain 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 a required height, based on aquifer parameters and expected recharge. Because groundwater migrates toward the closest drain the watertable forms planes of symmetry around each drain and the highest point of the watertable marks the groundwater boundary between them (Fig. 54). These boundaries define the groundwater catchment for each drain.

The extent of each plane and the depth and so the volume 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 in design to maintain the watertable at or below the desired height. 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 during design of the scheme 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. 54).

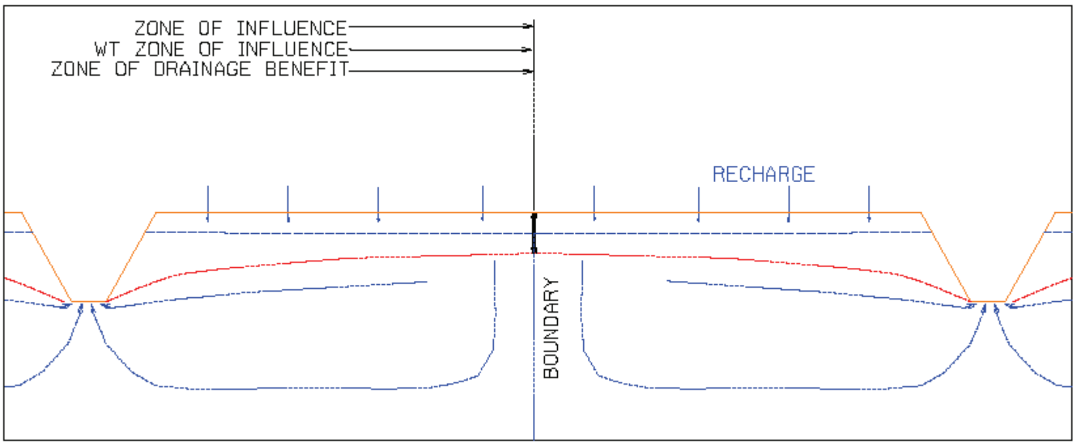

Figure 54 Watertable control by bounded drains

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 watertable response 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. 55). 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 (aquifer discharge (Appendix F)), 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 Morawa, cereal cropping. The extent of this adequate watertable control is the ZOB.

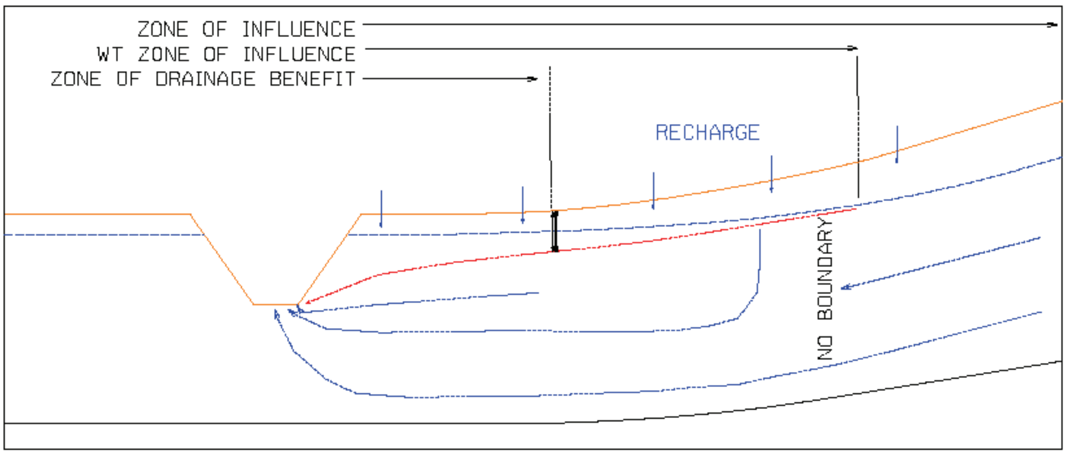

Figure 55 Watertable control near an unbounded drain

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 between drains. 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. The relationship between falling watertable height and reducing groundwater inflow to the drain exists regardless of whether the drain is the primary agent in controlling the watertable.

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 for an unknown duration. 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 quickly be 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 alongside. These are the drainage conditions that currently exist at Morawa.

Appendix B Discharge groundwater inflow and load

A salt mass balance was used to separate the components of drain discharge and basin recirculation from the total of pumped drain discharge from 22 December 2004 to 15 June 2005. The same approach was used to check the accuracy of the resultant water balance for the drain channel (Appendix CD 5.2).

A salinity of 59 536 mg/L was used in a salt mass balance to calculate the recirculated leakage from the basin into the cut-off drain. This salinity was equal to the average salinity of the leakage lost from the basin for the period December to June, above (Appendix CD 5.8). Using the salt mass balance approach separated 47 576 kL of basin recirculated leakage from the remaining 501 803 kL of drain discharge (Fig. 56) During the corresponding period, 2833 t of the total 20 785 t of salt was recirculated with the leakage water. The resultant discharge and salt load graphs still reflect high levels of distortion in response to the pumping frequency and duration. This is particularly evident in June 2005 when pumping was suspended pending completion of the contingency basin, and then re-started to remove the build-up of discharge from the sump and drain.

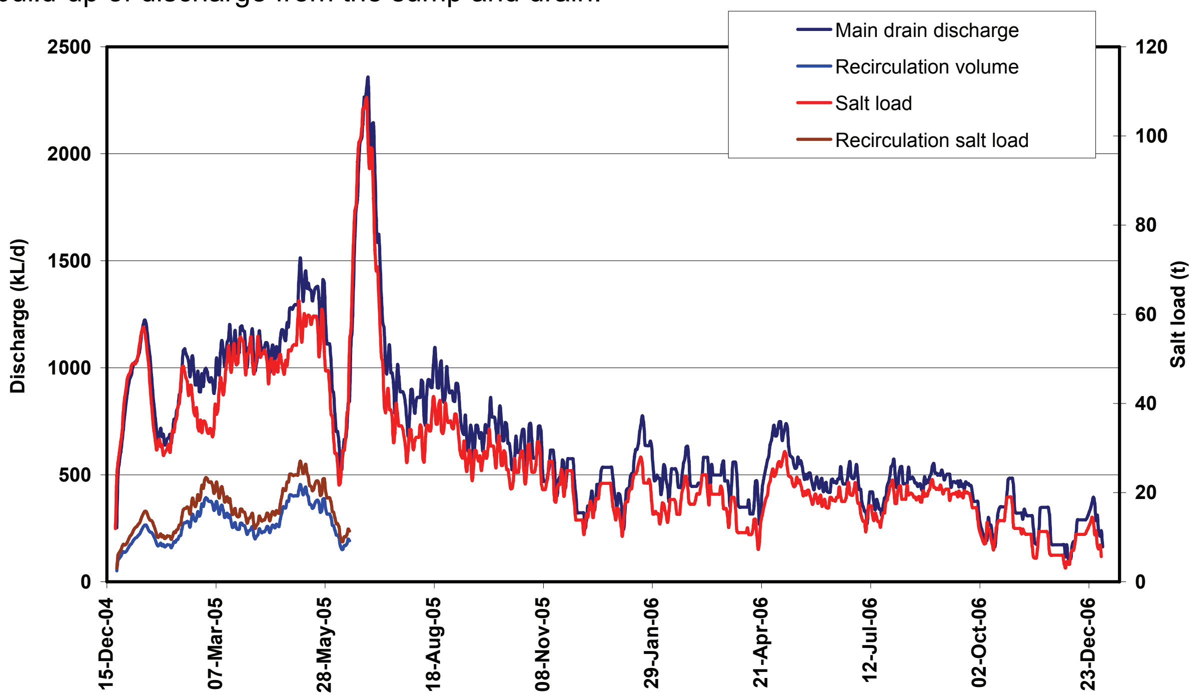

Figure 56 Total drain discharge and salt load less the recirculated basin leakage and salt load

Groundwater inflow (GWI) to the drain consisted of measured discharge plus evaporative loss to, and less rainfall and runoff from, the channel. Evaporative loss was calculated by applying evaporative-loss coefficients to the water in the channel and the lower 0.5 m of the drain batters (Appendix CD 5.2). The daily evaporation rate was reduced to compensate for the average daily salinity of the water in the drain (Turk 1970). Contributions to drain discharge from rainfall were calculated using rainfall runoff coefficients of 0.9 for the channel floor and 0.25 for the channel batters (Appendix CD 5.2).

The resultant groundwater inflow to the drain of 561 768 kL consisted of drain discharge plus evaporative losses and less rainfall contributions. To check the accuracy of the result, a salt mass balance was applied to the drain discharge using a uniform input salinity of 37 314 mg/L to represent the salinity of the groundwater inflow. This input salinity coincided with the average salinity of the shallow groundwaters beneath transects 1–3. The result was within 1% of the previously calculated groundwater inflow.

The calculated average salinity of the groundwater inflow to the drain was 35 246 mg/L, closely corresponding with the input salinity of 37 314 mg/L, above, used to check the calculations. The average daily flow-weighted salinity of the drain discharge was 40 404 mg/L for the same period. The final results show daily drain discharge, flow-weighed salinity, groundwater inflow and groundwater inflow salinity for the main drain, excluding the effects of basin leakage (Fig. 57). As expected, groundwater inflow exceeded drain discharge at all times and the difference between the salinity of flow-weighted discharge and of the groundwater inflow increased during the summer periods as salts were concentrated in the drain discharge by evaporation.

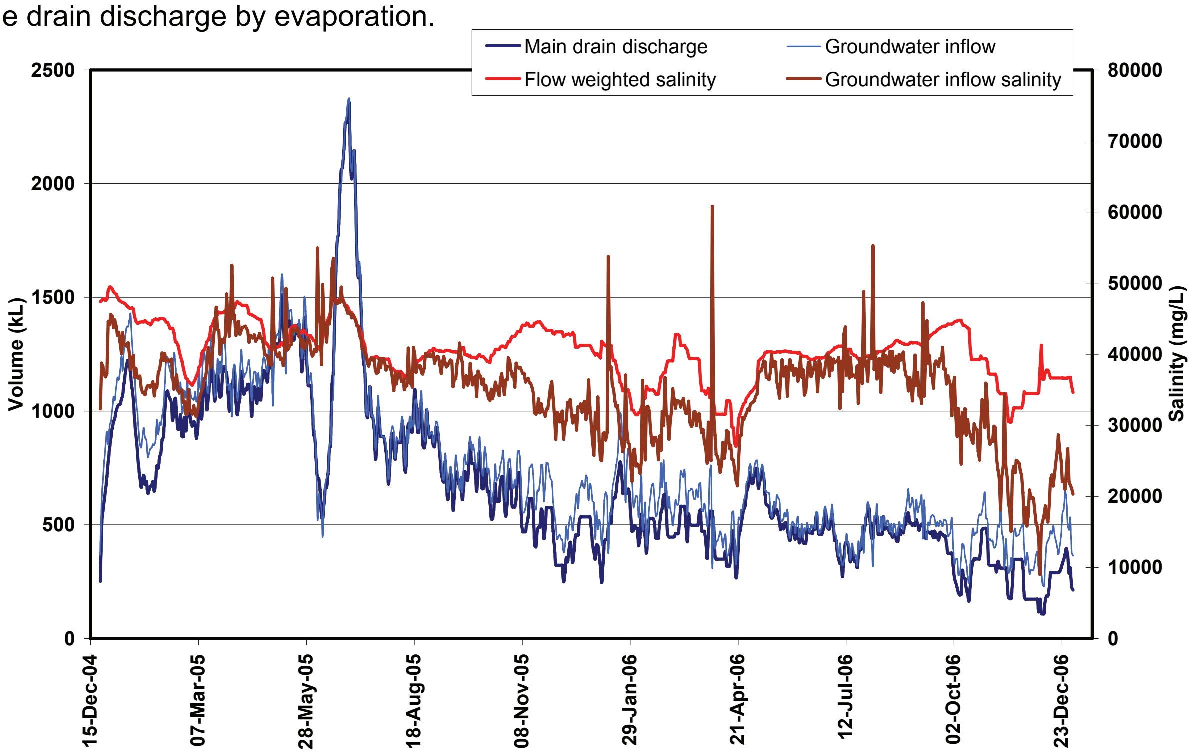

Figure 57 Total drain discharge and salinity compared to groundwater inflow

Appendix C Groundwater heads and drain discharge

Appendix A introduced the relationship between the watertable level or groundwater head alongside an effective drain and its corresponding groundwater inflow. Rising and falling watertables alongside the drain cause corresponding rises and falls in the groundwater head between the watertable and channel floor. If soil is sufficiently permeable the rising and falling head should be associated with increases and decreases in groundwater inflow to the drain.

Cox & Tetlow (in press) used the relationship between groundwater inflow to the drain and groundwater heads at various distances from the channel to explore the possible extent of the WT-ZOI of the drain. The rationale was that a high correlation between fluctuating groundwater head and inflow indicated that a high degree of control was exerted on the watertable by the drain. Typically, a high correlation would exist for the relationship between head and inflow for bores close to an effective drain. The correlation would steadily diminish with increasing distance from the drain as the watertable became progressively affected by response lags and other variables such as evaporation. The extent of the WT-ZOI was defined as the distance from the channel where there was no or a negative correlation between head and inflow.

The groundwater inflow hydrograph was amended to redistribute the trough and peak values (Fig. 57) to produce a more uniform hydrograph during 27 May–12 July 2005. Superimposing the watertable hydrographs for transect 3 on the groundwater inflow hydrograph (Appendix B) shows steadily increasing inflow corresponding with decreasing groundwater head from 22 December 2004 to 27 April 2005 (Fig. 58). From about 4 July 2005 groundwater heads and inflow show similar downward trends.

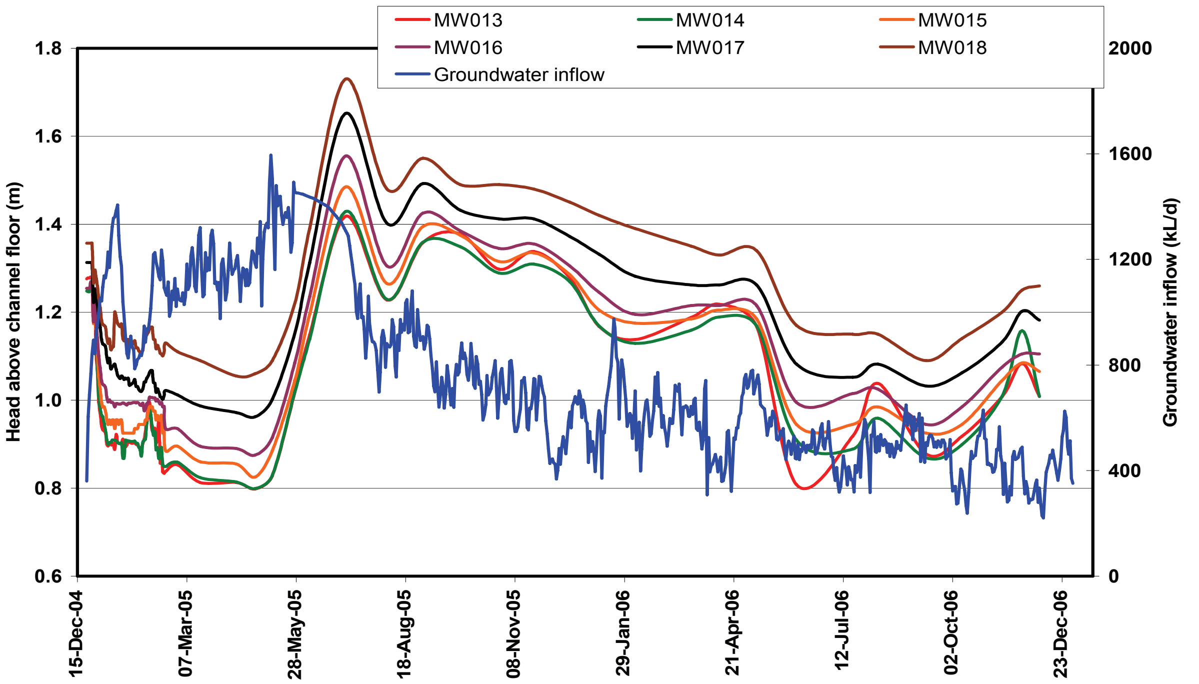

Figure 58 Watertable head above the channel floor for transect 3 and groundwater inflow

The relationship between groundwater heads and inflow at Morawa was explored using the watertable levels from transects 1–4 and the comparison bores (Section 4.1), and corresponding groundwater inflow. The closeness of the relationships was expressed as a correlation coefficient with a value of 1.0 representing a fully corresponding relationship, and 0 representing no relationship. Values from 0 to –1.0 represent inverse relationships between head and inflow, that is low groundwater heads associated with high inflows.

A moderately positive relationship existed between all of the groundwater heads in the watertable bores along transect 1 and groundwater inflow to the drain (Fig. 59). The remaining transects produced inconsistent or negative correlations that closely corresponded to those of the comparison bores. The inconsistent and negative results for transects 2–4 were assumed to be caused by the initial period of falling groundwater head and increasing discharge from 22 December 2004 to 27 April 2005 following construction of the drain (Fig. 59).

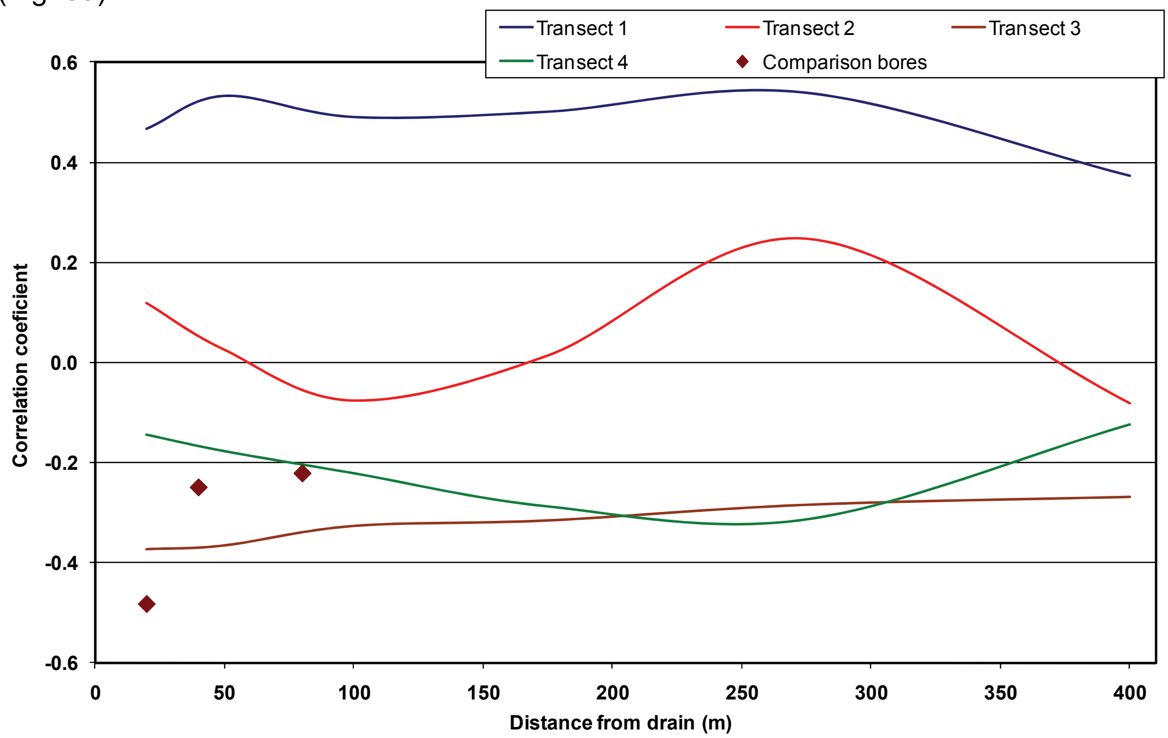

Figure 59 22 December 2004–6 December 2006 correlation coefficients for the head vs inflow at different distances from the drain

To test the assumption that the early responses of the groundwater levels and inflow to the drain were responsible for the negative correlations between the two, they were recalculated for the period 27 May 2005–6 December 2006. For all of the transect bores and comparison bores 031 and 032 moderate to high positive correlations were found and only comparison bore 030 retained a negative correlation(Fig. 60).

Only the bores in transect 1 produced the expected result of a high correlation between head and inflow close to the drain that diminished with increasing distance (Fig. 60). There was no change in the correlation of the bore 400 m from the drain when compared to the correlation for the same bore for the entire period of monitoring (Fig. 59).

If projected forward using a second-degree polynomial line, the transect 1 curve intersects the 0 correlation level at 500 m from the drain. Using the same approach for the other transect curves gives 0 correlations at 1000 m from the drain for transects 2–3, and no point of intersection for transect 4. The results from transects 1–3 tend to suggest a WT-ZOI of 500–1000 m for the drain.

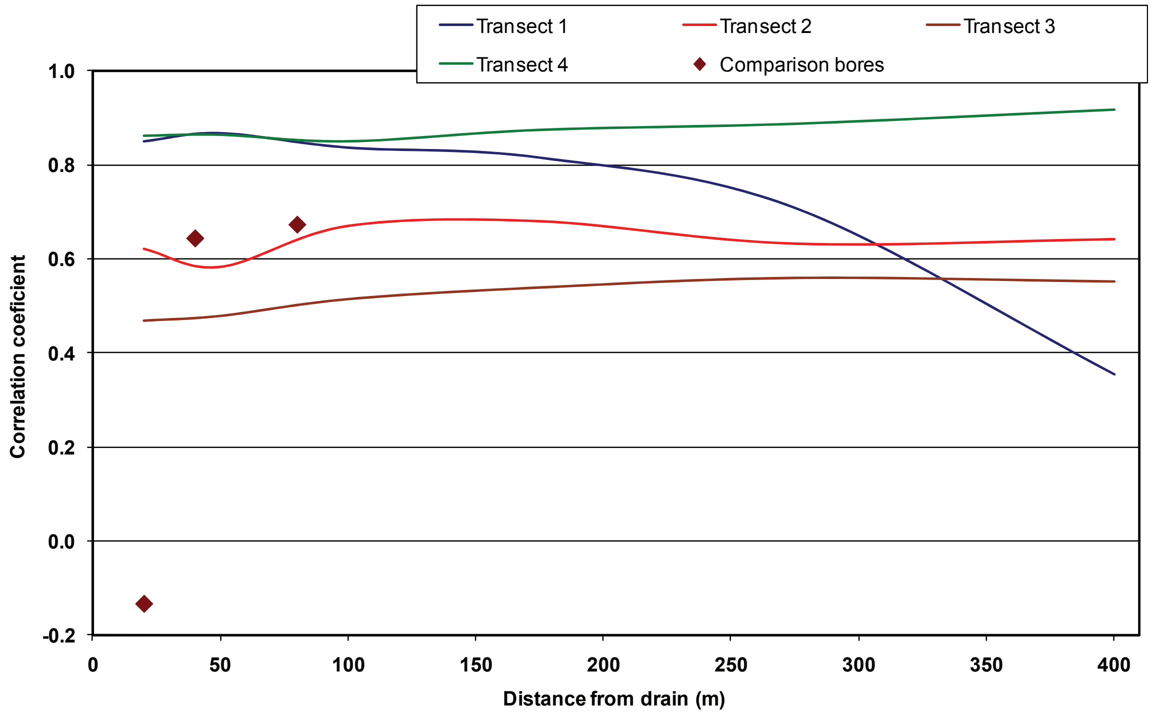

Figure 60 27 May 2005–6 December 2006 correlation coefficients for the head vs inflow at different distances from the drain

The uniform correlations between the groundwater heads along transects 2–4, and groundwater inflow to the drain have meant that the analysis results were inconclusive in terms of confirming the WT-ZOI of the drain. The moderate correlations between the groundwater heads in the comparison bores and inflow to the drain highlight the uncertainty surrounding the results.

The absence of a declining relationship between groundwater heads at increasing distance from the drain and groundwater inflows could lead to the following conclusions:

- The drain is controlling the watertable beyond the outer transect bores and encompasses the area surrounding the comparison bores 031 and 032.
- The drain is influencing the watertable and the aquifer beneath it but both inflows and watertable responses are largely affected by the same climatic variables.

Appendix D Groundwater storage change

Changes in aquifer storage beneath the drain and basin sites were calculated from the fluctuations in the watertable measured along transects 1–5. The aquifer storage changes were calculated from the change in cross-sectional area between watertable movements measured along each of the bore transects. The resultant change in cross-sectional area were multiplied by a specific yield value of 0.05 (Smedema & Rycroft 1983). This converted the changes in cross-sectional area to volumetric storage change beneath a 1 m wide section of the aquifer along each transect.

The removal of water from storage between one month and the next beneath transects 1–3 and 4 is represented by a negative value in the monthly time-step hydrograph (Fig. 61). The change in storage beneath the comparison bores was estimated by superimposing them along a 400 m transect, to produce results comparable to those of transects 1–4 (Appendix CD 5.4).

The monthly pre-drain change in storage along transects 1–3 (average) largely corresponds to that of the comparison bores (Fig. 61). From the start of monitoring in May 2004 to drain construction in December 2004, the average storage change along transects 1–3 increased by 2.0 kL, and by 1.6 kL for the comparison bores. After drain construction there were periods when the average storage beneath transects 1–3 increased far less than beneath the comparison bores (July 2005, May–June 2006). From drain construction to December 2006 the average storage change beneath transects 1–3 decreased 5.7 kL while beneath the comparison bores increased by 6.5 kL.

The rapid decline in storage beneath transect 4 to –4.3 kL in December 2004 was caused by pumping from the cut-off drain during the early stages of basin filling, prior to the onset of basin leakage. Storage increased by 9.5 kL from –4.3 to 5.2 kL by January 2005, in response to basin leakage and rising watertables (Fig. 61). Between the start and end of the monitoring period, average aquifer storage change beneath transects 1–3 reduced by 3.8 kL while increased by 18.3 kL, and 8.4 kL below transect 4 and comparison bores respectively.

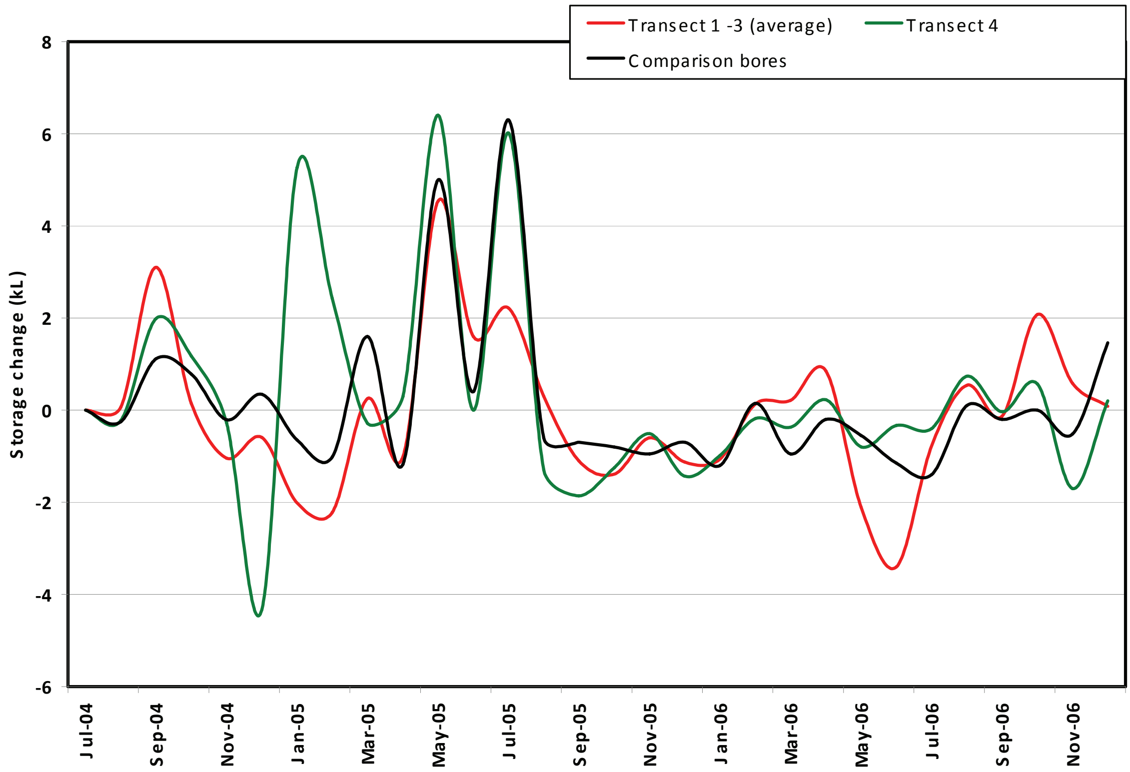

Figure 61 Aquifer storage changes beneath transects

The combined results from transects 1–3 were extrapolated to represent the aquifer storage changes beneath the 597 ha drainage site, and the results from transects 4–5 to represent storage change beneath the 60.5 ha basin site. At the commencement of drainage in December 2004–February 2005, groundwater inflow to the drain seemed to be derived mainly from groundwater released from aquifer storage. Inflow of 7868 kL during late December 2004 closely matched aquifer storage change of –8804 kL for the month (Fig. 62). Drain inflows of 32 974 kL and 29 681 kL also coincided with storage changes of –30 037 kL and –33 127 kL for the months January and February 2005.

The close relationship between inflow and storage change ceased in March 2005 with the onset of rainfall after the December 2004–February 2005 period of almost no recharge (Fig. 64). Groundwater inflow to the drain was 35 637 kL while aquifer storage increased by only 3744 kL during March 2005. During May 2005 aquifer storage increased by 67 800 kL while inflow to the drain was 41 351 kL (Fig. 62).

Although there was no significant net change in aquifer storage between the start and end of the monitoring period, this analysis shows that 277 343 kL of water was removed from the aquifer beneath the drainage site (Fig. 63). Rain-fed recharge and aquifer discharge intermittently refilled 249 178 kL of the drained aquifer storage, resulting in a net reduction of 28 165 kL, reflected by the final difference between the cumulative aquifer filling and draining results (Fig. 63).

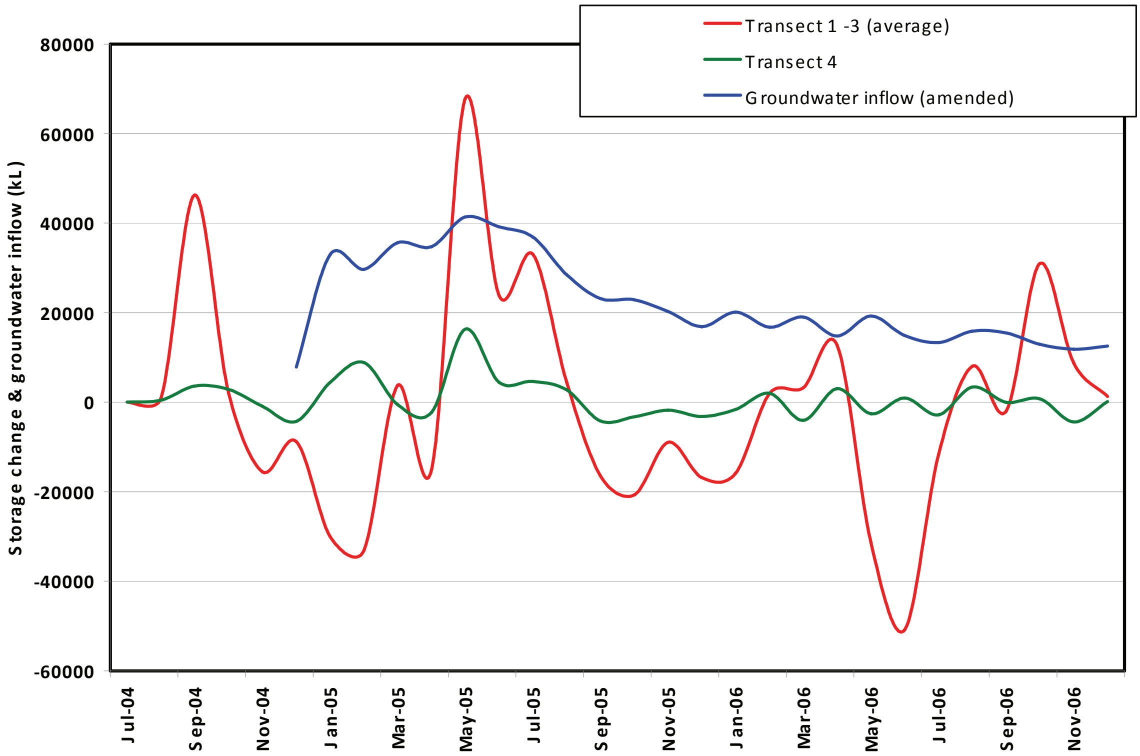

Figure 62 Aquifer storage changes beneath the drain and basin sites and groundwater inflow from Appendix C

There was a distinct separation between periods of net filling and draining of the aquifer beneath the drainage site. Aquifer storage was filled from the commencement of monitoring to October 2004, May–August 2005 and June–December 2006 (Fig. 63). Between these periods, the aquifer drained, with the largest volume of drainage of over 100 000 kL occurring during the period surrounding drain construction in December 2004.

The commencement of aquifer filling was seen to be in response to the onset of a significant rainfall event marking the beginning of the winter rainfall season. This was most evident in May 2005, where rapid filling of the aquifer was seen (Fig. 63) in response to May recharge (Fig. 64). Although refilling of the aquifer commenced in response to significant March 2006 rainfall, with the lack of follow-up rainfall and recharge the aquifer reverted to draining until July 2006.

Although all groundwater entering the drain had to pass through its surrounding aquifer, this does not mean that groundwater inflow to the drain must be reflected by changes in aquifer storage. Steadily moving groundwater can flow through the aquifer without change in storage. It is also unlikely that this storage change analysis captured all of the movements of groundwater into and out of the aquifer because of the infrequent watertable measurements in some bores. For these reasons, the drainage of water reflected by changes in aquifer storage represented less than half that of total groundwater inflow to the drain (Fig. 63).

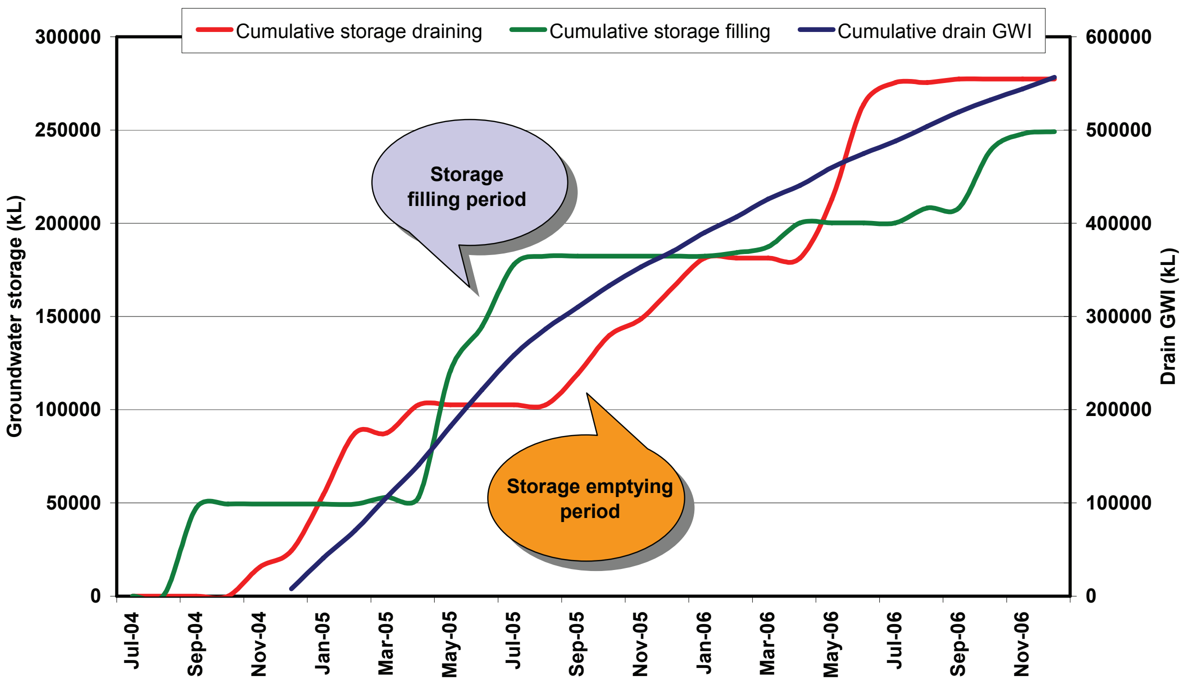

Figure 63 Cumulative monthly storage filling and drainage balances for transects 1–3

Appendix E Rain-fed recharge

An average total of 8.5% rain-fed recharge from 577 mm of rainfall was calculated from watertable fluctuations beneath the drainage and basin sites (Appendix CD 5.5). During the post-drainage to December 2006 period, 34.4 mm of recharge was equivalent to 8.4% of the 410 mm of rainfall. Although 2006 rainfall was 62% of 2005, it produced an equivalent of 72% (14.2 mm) of 2005 recharge. These results highlight that the relationship between rainfall and recharge is not proportional, and in reality is affected by many variables.

Rain-fed recharge for the water balances was estimated using the relationship between rainfall and short-term (post 2 day) groundwater level rise (Appendix CD 5.5). Bores with near-continuous water level measurements (002, 008, 014, 020 027 and 032) were used to develop rainfall–recharge exponential relationships (Fig. 64), and these equations applied back to the rainfall to produce more consistent simulated rainfall–recharge responses between the bores.

The individual simulated monthly recharge results for each bore were within +-80% of the average of all of the bores, excluding bore 014. The values from 014 were constantly 1.5–2 times higher than the next highest value for the other bores. This possibly reflected additional recharge generated from the inundation associated with the poor surface drainage surrounding this bore in transect 3. Water-level interference between the drain and the bore could also have affected the initial rainfall–watertable reduction. As the cause is not clear, the values have been used in the calculation of total recharge to the site to represent the higher potential for recharge on the parts of the valley floor subject to inundation.

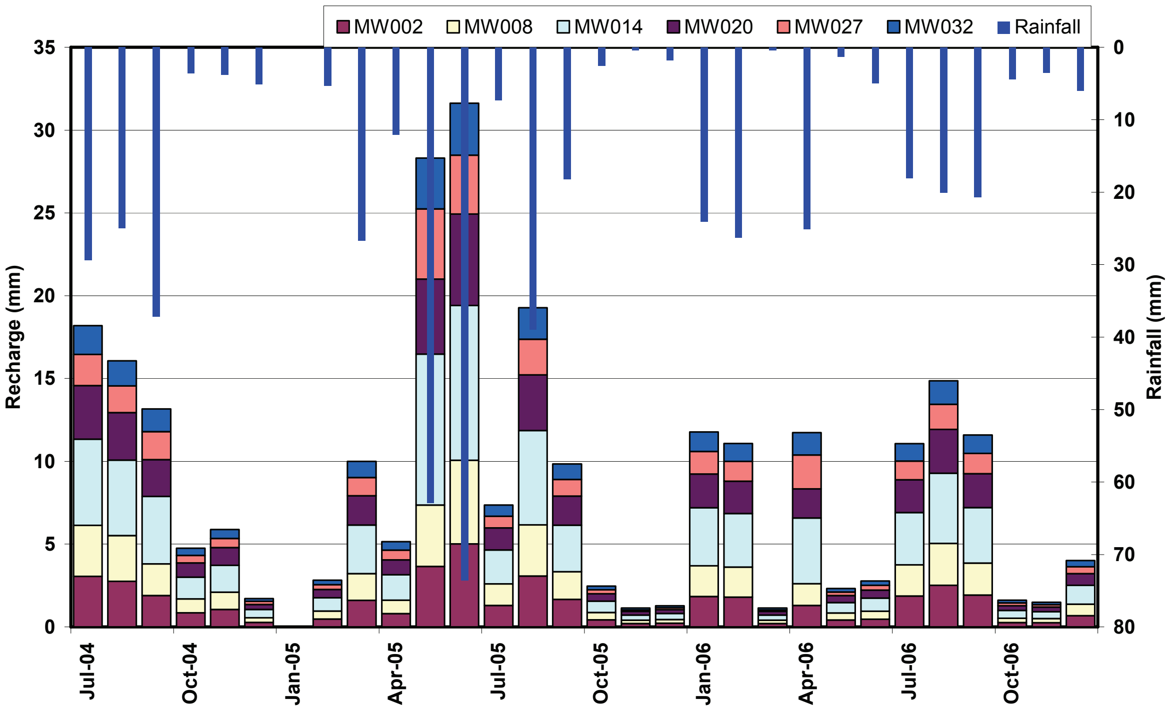

Figure 64 The relationships between rainfall and calculated recharge for the six bores in the analysis

Recharge to the drainage site was represented by the average results of bores 002–014. For the entire period of monitoring, rain-fed recharge was 54.0 mm (10.6% of rainfall) with a monthly range of 0–6.5 mm (0–60% of corresponding monthly rainfall). Recharge to the basin site represented by the results from bore 020 was 45.9 mm (9.0% of rainfall) with a monthly range of 0–4.2 mm (0–52% of corresponding monthly rainfall).

Appendix F Aquifer discharge

The aquifer beneath the valley floor appears to be well connected to the aquifer beneath the hillsides and reinforces the conclusion that much of the groundwater in the valley floor could come from hillside and surrounding aquifer discharge. In the context of the unbounded drain (Appendix A) 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 Morawa drain.

Aquifer discharge can appear to move large distances towards a drain because the flow paths are not only horizontal but radial. Radial groundwater flow can occur beneath the Morawa site because the ratio of aquifer thickness to drain depth is high (approximately 15:1), the drain is unbounded and the aquifer unconfined (Ritzema 1994). The flow-net model demonstrates that 'groundwater streamlines' develop extensive radial flow pathways towards the drain (Fig. 65). The flow lines can penetrate the full thickness of the aquifer and greatly enhance the lateral impact of the drain.

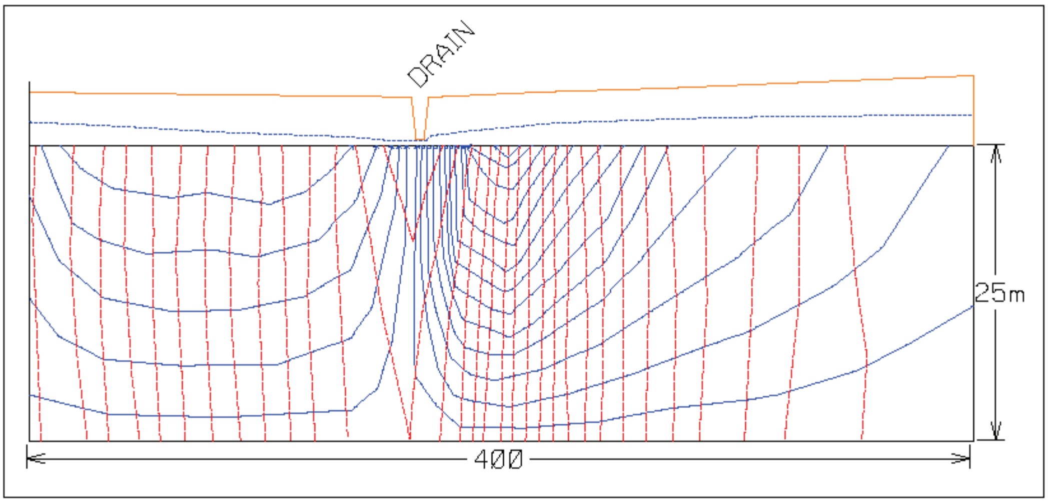

Figure 65 Flow net through part of transect 3 to bore 046

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

An alternative approach is to calculate aquifer discharge at a chosen point 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: at 400 m for the Morawa drain (Section 5.1).

Put quite simply, the aquifer discharge contribution to the drainage site is calculated from its discharge or outlet end through a porous boundary rather than its recharge or inlet end elsewhere in the catchment. The method is based on applying the Darcy assumption (Ritzema 1994) to the slope of the watertable at a chosen point. For Morawa the representative points for the slope of the watertable were the outer transect bores, the slope being the hydraulic gradient between this and the next closest bore in each transect (Appendix CD 5.6).

The use of this method represents aquifer discharge as the relationship between groundwater heads and supply outside the drainage site compared to the groundwater heads inside the drainage site. This is believed to be an accurate reflection of part of the process that controls aquifer discharge from hill-slope to valley floor aquifer.

Before drainage, the discharge rate to the valley floor was partially regulated in the short term by the head differences between the valley floor and hillside aquifers. Localised recharge on the valley floor caused groundwater rise which in turn reduced aquifer discharge. As the watertable on the valley floor fell, the head reduction resulted in increased aquifer discharge.

With drain construction the watertable started to fall, first close to the drain and then further away. As soon as the depression in the watertable reached the arbitrary boundary, aquifer discharge increased. This increase appears to be a response to the lower groundwater heads and steeper hydraulic gradients created by drainage of the valley floor.

The drain appeared to have a significant impact on lowering the watertable along transect 3 by about 0.3 m during the two weeks after drain construction (Fig. 66). During this time aquifer discharge also reduced. After two weeks, the falling watertable response to the drain stopped in response to the halt in the declining aquifer discharge.

From early February to April aquifer discharge remained above its pre-drainage levels and the watertable along transect 3 continued its steady decline by a further 0.1 m. The remainder of the watertable and aquifer discharge hydrographs reflected a generally inverse relationship between the heads inside and outside the drainage site.

The hydrographs reveal what was observed in the field: the valley floor watertable close to the foot slopes did not start to reduce until the foot slopes and valley flanks appeared to have somewhat dried out. This drying out was apparent by the cessation of some of the intermittent seepage that occurred at some of the foot slopes. The drain did not start lowering the watertable until the supply of groundwater from aquifer discharge was reduced. The implications are that, at Morawa, aquifer discharge is an almost constant supply of groundwater that will sustain drain discharge in the absence of any other sources of groundwater to the drainage site.

Aquifer discharge into the drainage site (Fig. 44) was represented by the average aquifer discharge estimated for transects 1–3, through the drainage site perimeter of 15 200 m (Section 5.1). The aquifer thicknesses used in the Dupuit assumptions were based on the depth to basement identified from the drilling of the deep bores within each transect (Appendix CD 2.5a) and an aquifer hydraulic conductivity of 1.2 m/d (Section 2.5).

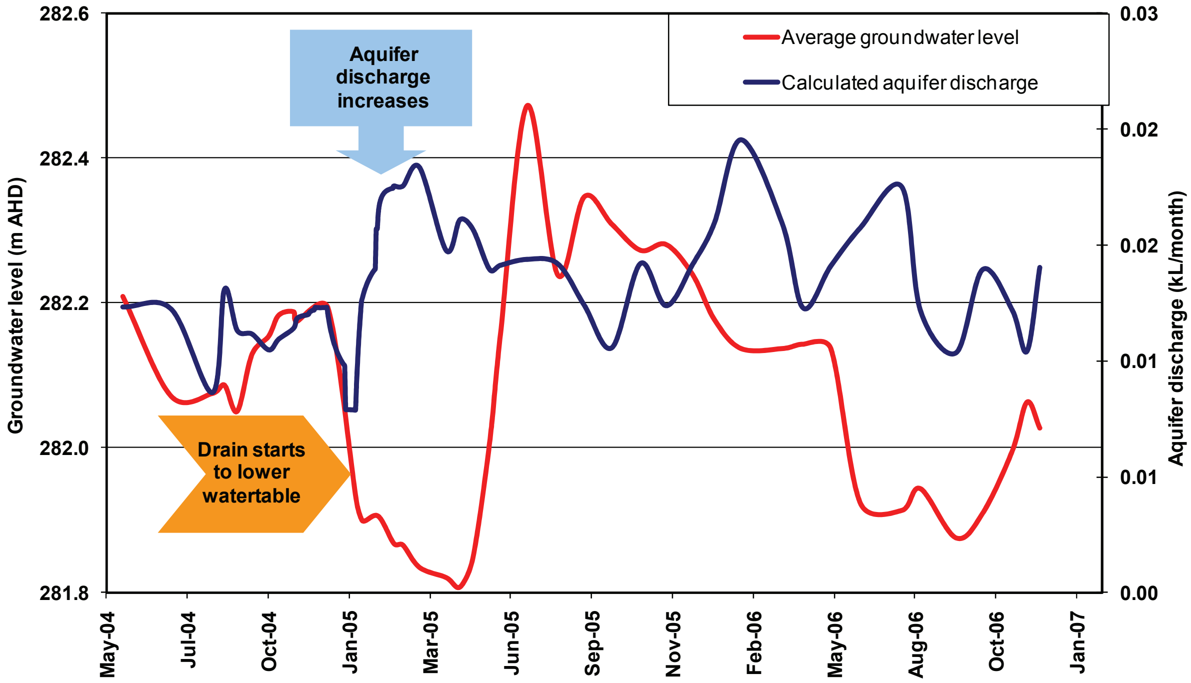

Figure 66 The inverse relationship between transect 3 watertable height and aquifer discharge

The results for transect 4 showed that aquifer discharge into the basin site changed to net aquifer discharge out of the basin site in response to basin leakage. The results from transect 4 were used to estimate aquifer discharge through the 2600 m perimeter of the basin site (Fig. 43) using the same approach and variables as for the drainage site.

Aquifer discharge into the drainage site exceeded groundwater from rainfall recharge by 172% during the post-drain monitoring period (Appendix CD 5.6). The 429 422 kL of post-drain aquifer discharge was equivalent to 77% of the total groundwater inflow to the drain, showing the potential for drain-flow to be dominated by aquifer discharge. In the five months prior to drainage, 82 684 kL of aquifer discharge went into the drainage site. This raises the question: what would have happened to this aquifer discharge before drainage?

Appendix G Capillary rise and discharge

Capillary rise and evaporative loss had the greatest *potential* to remove groundwater from the drainage and evaporation basin sites at Morawa. Capillary rise had the potential to transport about 1 300 000 kL of groundwater from the watertable to the soil surface within the drainage site during the monitoring period. Potential evaporative loss from the site was about 38 500 000 kL over the same time span.

Capillary rise is the lifting of a water column by surface tension between water and circumference of the capillary (Smedema 1983). The extent of capillary rise of groundwater above the watertable depends on the water tension in the root zone or soil surface, and the diameter and connectivity of the soil pores (capillaries). The upper extent of the capillary rise is called the capillary fringe.

As a watertable approaches the soil surface the capillary fringe enters the root zone, making the groundwater available for plant use. As the watertable keeps rising the capillary fringe reaches the soil surface and water can be lost by direct evaporation. The loss of groundwater by evapotranspiration or directly by soil surface evaporation is capillary discharge. The closer the watertable to the soil surface, the higher the potential capillary discharge.

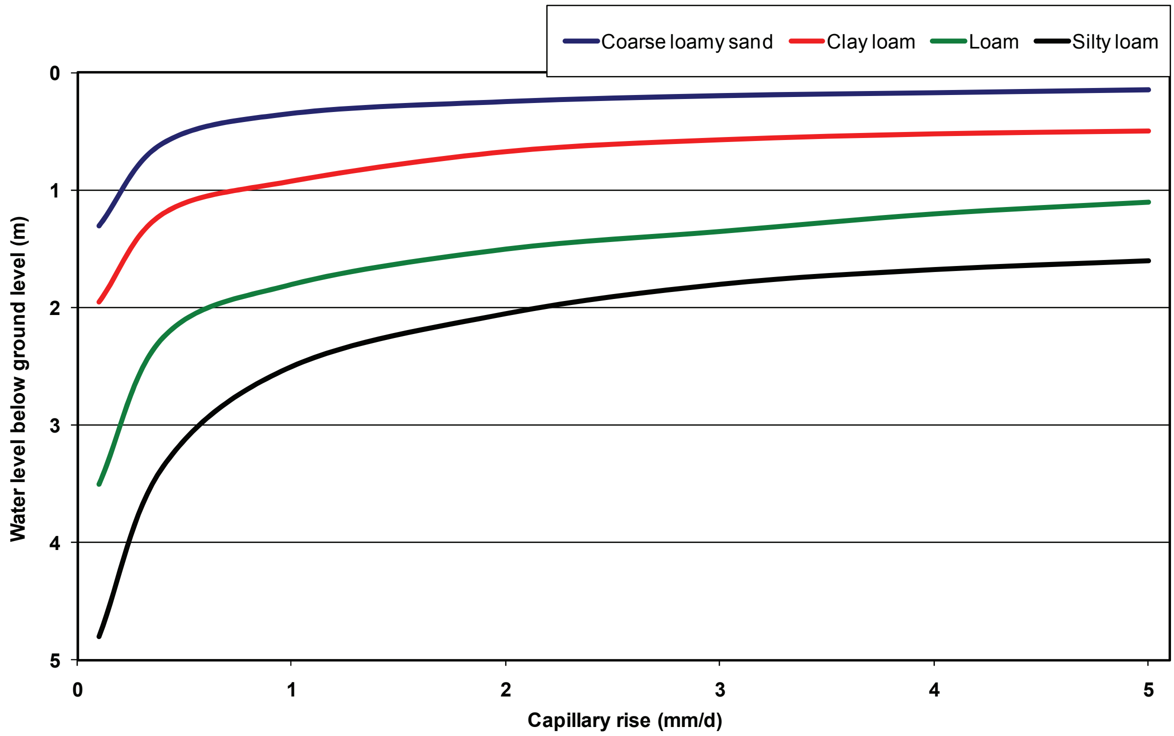

Figure 67 Rates of capillary rise with various soil types and depths to watertable

Source: Redrawn from Smedema & Rycroft 1983

The relationship between soil texture, watertable height and capillary rise has the 'Y' axis as the depth to the watertable and the 'X' axis as the rate of capillary rise (Fig. 67). The water tension is equivalent to a soil moisture pressure of 16 bar at the surface, the wilting point for most crop plants. For the silty loam soil type, capillary rise is near zero while the watertable is 4.8 m below ground level. As the watertable rises, capillary rise increases, and is 5 mm/d when the watertable is 1.6 m below ground level. In the other three coarser textured soils the watertable can be closer to the soil surface before having the same rates of capillary rise as the silty loam.

Capillary rise and discharge were believed to play significant roles in controlling both the pre- and post-drain watertable heights and groundwater losses across the Morawa project site. The watertable fell in response to capillary discharge as groundwater was removed and transported to the soil surface by capillary rise. The high rates of capillary discharge were substantiated by salt accumulating across the valley floor and on foot slope soil surfaces. As saline groundwater moved to the surface and evaporated, salts accumulated on the surface and in the root zone evident by a white crust on the land surface (Fig. 68).

Figure 68 Salt accumulated on the surface when the saline watertable was 1.2 m deep in this case

The estimated rates of potential capillary rise were calculated from the fluctuating watertable depths along transects 1–5 and for the comparison bores using a coarse loamy sand representative soil type (Fig. 67). The results were further extrapolated to represent potential capillary discharge for the drain and basin sites. If the calculated rate of capillary rise exceeded evaporative demand or if evaporative demand was satisfied by rainfall, capillary discharge was reduced to equal the evaporative demand or deficiency (Appendix CD 5.7), to simulate the effect of reduced soil moisture pressure at the soil surface reducing the rate of capillary discharge.

The relationship between watertable height and capillary discharge was evident within the drain and basin sites, particularly as the watertable rose nearer to the soil surface (Fig. 69). Mostly, fluctuations in the watertable below 0.8 m below ground level resulted in only minor fluctuations in capillary discharge. As the basin site watertable rose above 0.8 m, the capillary discharge rate rose significantly and, at 0.36 m deep, the potential capillary discharge was 430 kL/ha in June 2005 (Fig 69).

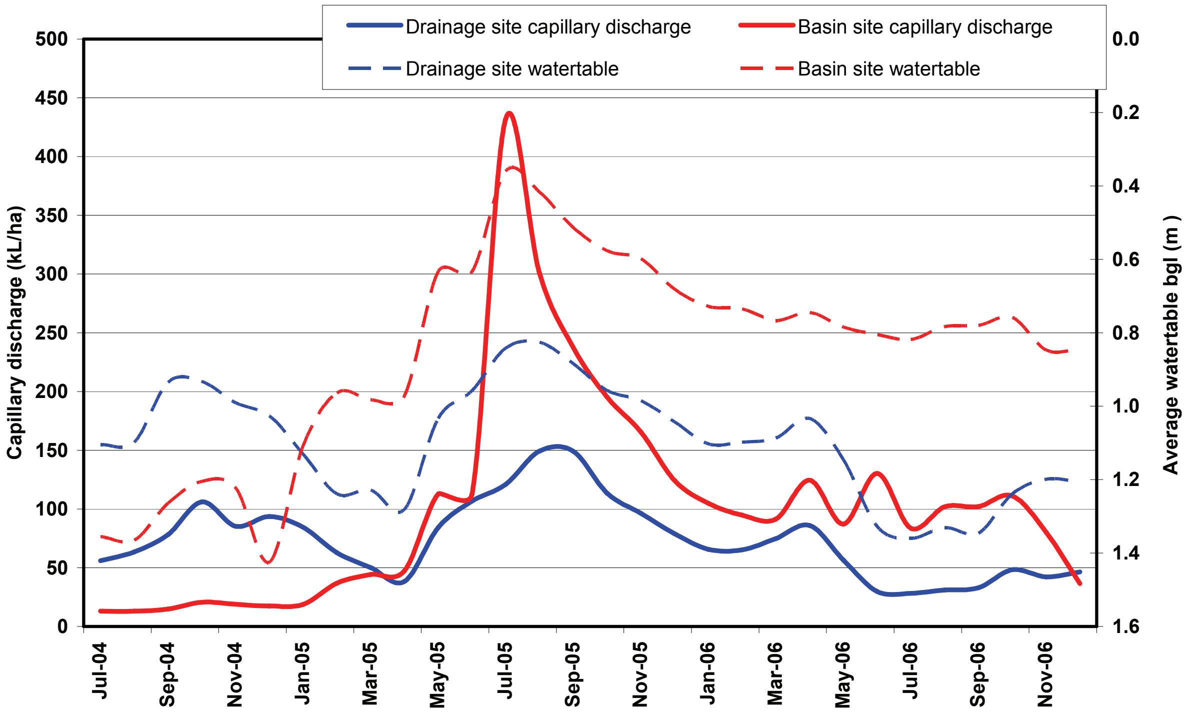

Figure 69 Monthly capillary discharge rates and average depths to watertable for the drain and basin sites

The total potential loss of groundwater by capillary discharge from the drainage site was equivalent to 2225 kL/ha for July 2004–December 2006. The average watertable depth for the same period was 1.10 m and total evaporative potential equivalent to 64 567 kL/ha. Expressed in millimetres depth across the site, 44.3 of the 222.5 mm of capillary discharge occurred in the pre-drain six month period, and 178.2 mm in the post-drain 24 month monitoring period. The results show there was no significant change in the pre- and post-drain 7.4 mm average monthly capillary discharge.

With an average depth to watertable beneath the basin site of 0.86 m, the total potential capillary discharge was 3074 kL/ha (307.4 mm). The average capillary discharge from the basin site was about 16 mm/month whilst the pre-basin watertable was 1.3 m bgl. For the 24 months post-basin, average monthly potential capillary discharge was 124 mm while the watertable was about 0.7 m bgl.

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)
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
Bedrock	The solid rock underlying the regolith
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 of the components of a drain: channel, berms and levees (if present)
Drainage catchment	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
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/fall recharge	Recharge from the percolation of rainfall and runoff to the groundwater system (mm)
Recharge	The addition of water to the groundwater system (mm)
Regolith	The layer of noncohesive or cohesive soil and rock material of whatever origin that nearly everywhere covers the surface of the land and is above the bedrock
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 water 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	(ųr) 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
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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