Water balance and salinity trend, Toolibin catchment, Western Australia

DEPARTMENT OF Conservation AND LAND MANAGEMENT

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1. Introduction

1.1 Background and objectives

Human-induced salinisation of the Australian continent's land and water has been occurring for last the 100 years, with some early reported incidents occurring in the Western Australian Wheatbelt (Bleazby, 1917; Wood, 1924). It is well documented that natural salinisation of parts of the Australian continent is due to climate throughout the geological record, (Bowler, 1990) as is demonstrated by the salt lakes of southern Australia. It is also well documented, however, that the recent increase in land and surface water salinisation occurred because native vegetation was cleared (Peck & Hurle, 1973). The changes in land use have altered the natural water balance, and increased recharge by more than one order of magnitude (Peck 1978; Allison & Hughes, 1978). The increased recharge rates have caused a substantial rise in groundwater levels, resulting in land salinisation and seepage into the surface water systems.

Toolibin Lake and its catchment (48,000 hectares) are affected by dry land salinity, similar to most of the valley floors and low-lying areas (McFarlane et al., 1989). The salinity of Toolibin Lake has increased, due to the rise in salinity of the North Arthur River (that drains approximately 90% of the catchment), and the level rise of saline water under the lake. These factors have resulted in a deterioration of the lake's ecosystem (George and Dogramaci, 2000).

Toolibin Lake is one of a chain of lakes occupying a palaeodrainage valley that forms part of the Northern Arthur River System (Figure 1). Toolibin Lake (and the surrounding reserves) is an important breeding habitat for native flora and fauna and has been listed under the Ramsar Convention, as a Wetland of International Importance.

Although there have been numerous regional and catchment-scale investigations into the groundwater dynamics and surface-water hydrology, few studies to date have addressed the interaction of surface and groundwater, and its ultimate impact on the ecology of Toolibin Lake. The objectives of this review are to collate available data on the Toolibin catchment, describe groundwater dynamics and related interaction with surface water, and present a model of the catchment in the light of recent investigations. In addition, a salt and water balance of Toolibin Lake has been prepared and modelling undertaken to review the impact of several management options.

The hydrological processes associated with dry land salinity and the salinisation of surface water are described in this report using data compiled over the last 20 years. A description of the landscape's evolution and its impact on the development of dry land salinity is presented. The effects of treatment options to control and reduce the impact of salinity are explained. Possible variations in salt load and associated stream flow changes are estimated for a range of treatments.

1.2 The impact of regional geology and climate on groundwater flow

The geological evolution of the Australian continent – particularly of the Great Western Shield that encompasses most of Western Australia – has had a major influence on the hydrogeology, groundwater flow, and consequently, the development of dry land salinity. Understanding the history and evolution of the landscape therefore, is a prerequisite to understanding the processes that have caused the rapid development of dry land salinity and the deterioration of ecological systems over the last 100 years.

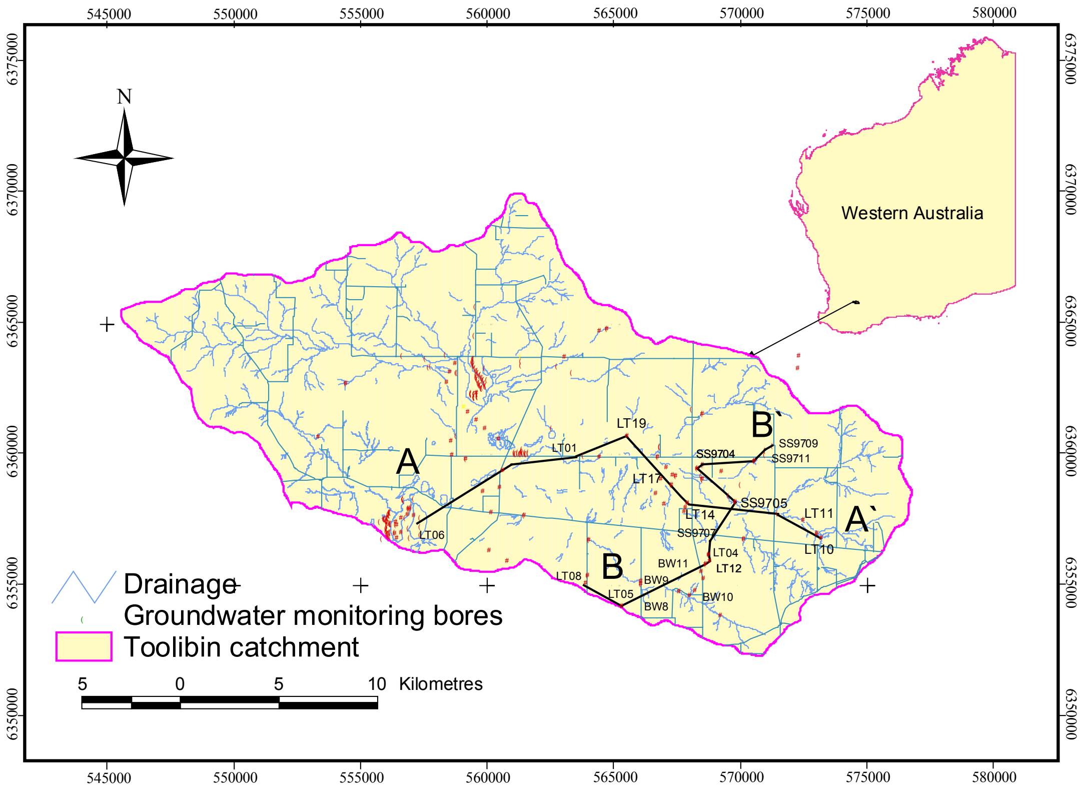

Figure 1. Toolibin catchment location map and the two hydrogeological transects.

The Toolibin catchment is located on the Yilgarn Craton, a giant, ancient and relatively flat crustal unit that has been stable for the past 2,400 million years. During this period the area has been little deformed by comparison with adjacent parts of the crust (GSWA, 1990). For much of this geological time, the Craton has been a part of the Gondwana super continent.

Prior to the continental rift (separation of Gondwana from Antarctica and the formation of the Australian continent) ~ 1,600 million years ago, the Craton was covered by ice and glaciated. Glaciation of this part of Gondwana may have caused the erosion and created a relatively flat landscape with river systems characterised by low gradients. The current palaeodrainage pattern in the Southwest region had developed throughout the Yilgarn by the late Cretaceous, 65 million years ago (Beard, 1999).

A further continental rift 45 million years ago during Tertiary periods, and the consequent rise of the Yilgarn Craton, changed and in many cases reversed the drainage systems. Uplift along the Darling Range associated with further continental movement blocked rivers, allowed sediments to accumulate and large palaeo-lakes to form.

The existing drainage system of the south-western part of Western Australia follows two main patterns, west south-westerly and north north-easterly. The latter is the predominant lineament pattern in the south-western part of the Western Shield of the Yilgarn Craton. These drainage systems have been active since the Eocene, with open systems discharging into the Indian Ocean.

The continued uplift of the Yilgarn Craton may have caused the damming of the rivers flowing in a westerly direction towards the sea and the formation of internal drainage systems by the recapture of streams and the formation of inland lakes. These lakes would be restricted to the major drainage lines during dry periods and would expand to flood large areas during wet periods. The continuous flow of these palaeodrainage rivers and transport of the weathered material from the upper part of the catchments, led to the formation of the sedimentary sequence found in the valley flats of the Toolibin catchment.

In most palaeodrainage systems there is a common lithostrato graphic depositional pattern; alluvial sands and gravels with lignitic beds at the base trending upwards into lacustrine clays and finally into modern alluvial and aeolian sands and clays. This period of sediment accumulation (Eocene) was also marked by a lack of sediment from the geological record in the Perth Basin. Dated lignitic materials in these channel systems suggest that they contain both Eocene and Pliocene sediments, indicating two different periods of sedimentation (Salama, 1994).

Sediments in the lake and river from various parts of the continent show that over the last half a million years, there have been dramatic changes in climate that have had a profound effect on river flow, lake formation and the movement of salt across the landscape. Furthermore, the modern climatic gradients that have developed over most of the south-west have promoted the accumulation of salt, brought into the landscape by rainfall. The steep decline in rainfall from west and southwest to east with a corresponding increase in evaporative loss means that most of the rain falling on the catchment was either taken up by vegetation or lost to the atmosphere by evaporation. The native vegetation transpired the small amount of rainfall that infiltrated the soil. However, salts entering the soil with rainfall are excluded by plants and therefore accumulate in the soil profile.

Prior to the vegetation being cleared, there was little or no recharge and hence no seepage of deep groundwater to the surface, so the surface runoff generated in south-west catchments was much fresher than at present. The discharge of this relatively fresh runoff in the years with higher than average rainfall (less than 1:20 years) was used to fill lakes Toolibin and Taarblin, providing an ideal freshwater environment for the diverse fauna and flora. While high salinity levels have caused most of these species to disappear from Taarblin, many still persist in Toolibin Lake.

Post clearing, the water balance of the Wheatbelt catchments changed with increased input from rainfall, which resulted in the rise of the watertable and eventual seepage of saline groundwater, and subsequent evaporation, into the low-lying areas of the catchment. The mixing of saline seepage and runoff, and the ultimate salinisation of the main drainage lines and creeks, resulted in the degradation of the ecosystems within discharge areas including wetlands. Previous investigations carried out in various catchments and wetlands suggest that the salinity of the inflow into lakes might have increased by more than four fold in the last 50 years (P.W.D., 1984).

1.3 The impact of topography on groundwater flow

Toolibin Lake occupies broad valley flats that represent the ancient drainage system containing water borne deposits of sands and clays (George and Dogramaci, 2000). The fluvial plain at Toolibin Lake is about three kilometres wide and is bounded on the east by aeolian dune deposits, which overlie fluviatile sediments and in situ weathered granite (Dogramaci, 2000). The western part of the plain merges with weathered basement and is overlain by thin colluvium. Toolibin Lake is located at the boundary between these systems, with its eastern flanks overlying lacustrine sediments, and to the west, alluvial sequences. There is a palaeochannel system 300 metres wide and 40 metres deep in the middle of the eastern sequence (beneath the lake), extending approximately five kilometres in a north-westerly direction (De Silva, 1999). The well-vegetated dunes along the eastern edge of the lake lie over lacustrine sediments (covering the once large palaeolake) and have a greater infiltration rate due to their sandy soils.

The elevation of the Toolibin catchment ranges from about 360 metres Australian Height Datum (AHD) at the north-eastern boundary of the catchment to 298 metres AHD at Toolibin Lake (Figure 2). The eastern part of the catchment has gently undulating flats that are characteristic of the ancient landscape of the Blackwood River Basin (Bettenay and Mulchay, 1972). The western part on the other hand is more undulating and characterised by a relatively well-defined drainage system that is similar to the rejuvenated landscape of the Blackwood River Basin (Bettenay and Mulchay, 1972).

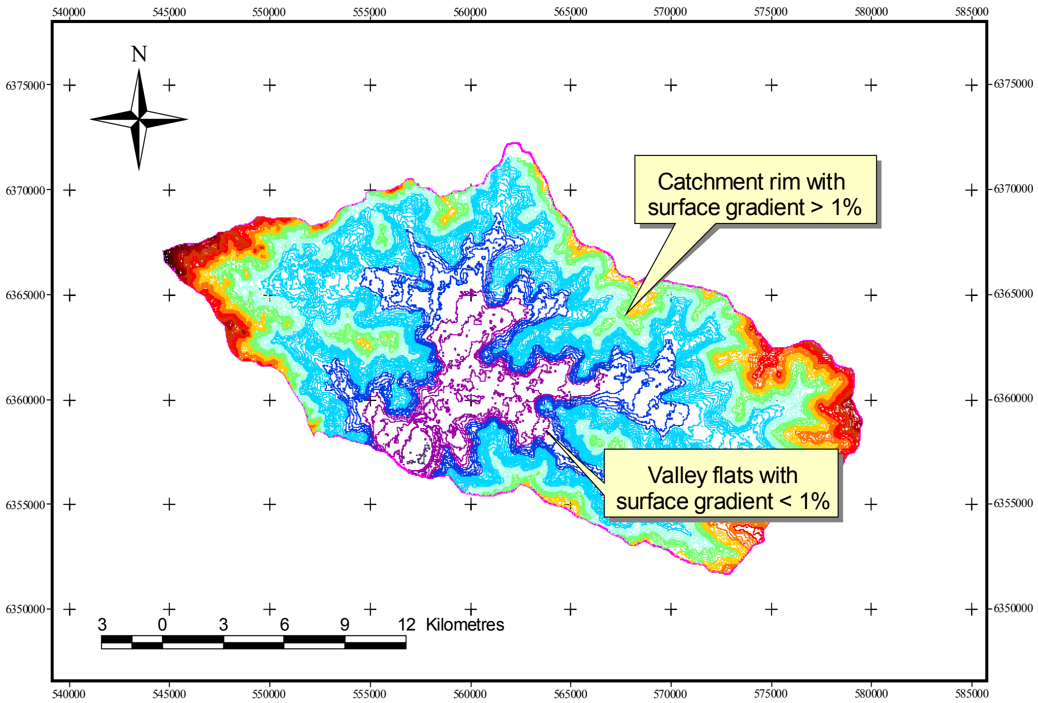

Figure 2. Topography of the Toolibin catchment. The elevation ranges from about 360 AHD at the eastern catchment boundary to 299 AHD at Toolibin Lake.

Based on the gradient, the catchment can be divided into two areas; the first area with gradient less than 1% occurs at the catchment rim and comprises about 65% of the catchment. The second area with a gradient more than 1% (valley flats) occurs along the main drainage line and comprises about 35% of the catchment (Figure 3). These topographical characteristics combined with the micro elevation within the valley flats may have an important control on the local (hillslope scale) and intermediate (catchment and inter-catchment scaled) groundwater flow systems.

Generally, the watertable in the areas characterised by low gradient (valley flats), conforms to the surface topography. The groundwater flow system in this type of terrain may be controlled by a variably connected, local flow system (i.e. recharge occurs in nearby uplands and flows laterally toward the valley flats, or occurs within the flats). The micro elevations within the valley flats produce numerous sub-systems within these local flow systems. Conceptually, water that enters the flow system in a given recharge area will be discharged at the nearest topographical low or it may be transmitted to the nearest discharge area. As the ratio of the depth to lateral extent of an aquifer becomes smaller, the groundwater flow systems are more likely to be controlled by local rather than intermediate flow systems. This is the most likely scenario for the groundwater flow systems in the valley flats as demonstrated by McFarlane et al, (1989).

Conversely, even in poorly-undulating terrain such as occurs in the valley flats, the presence of basal units that are characterised by a relatively high hydraulic conductivity creates a preferred pathway for flow that passes under the overlying local systems. The existence of highly transmissive material may allow intermediate systems to develop even in areas of marked local relief. In the

Toolibin catchment, palaeochannel sediments create this opportunity, as may extensive saprock aquifer units in which lateral flow is unaffected by dykes or similar features.

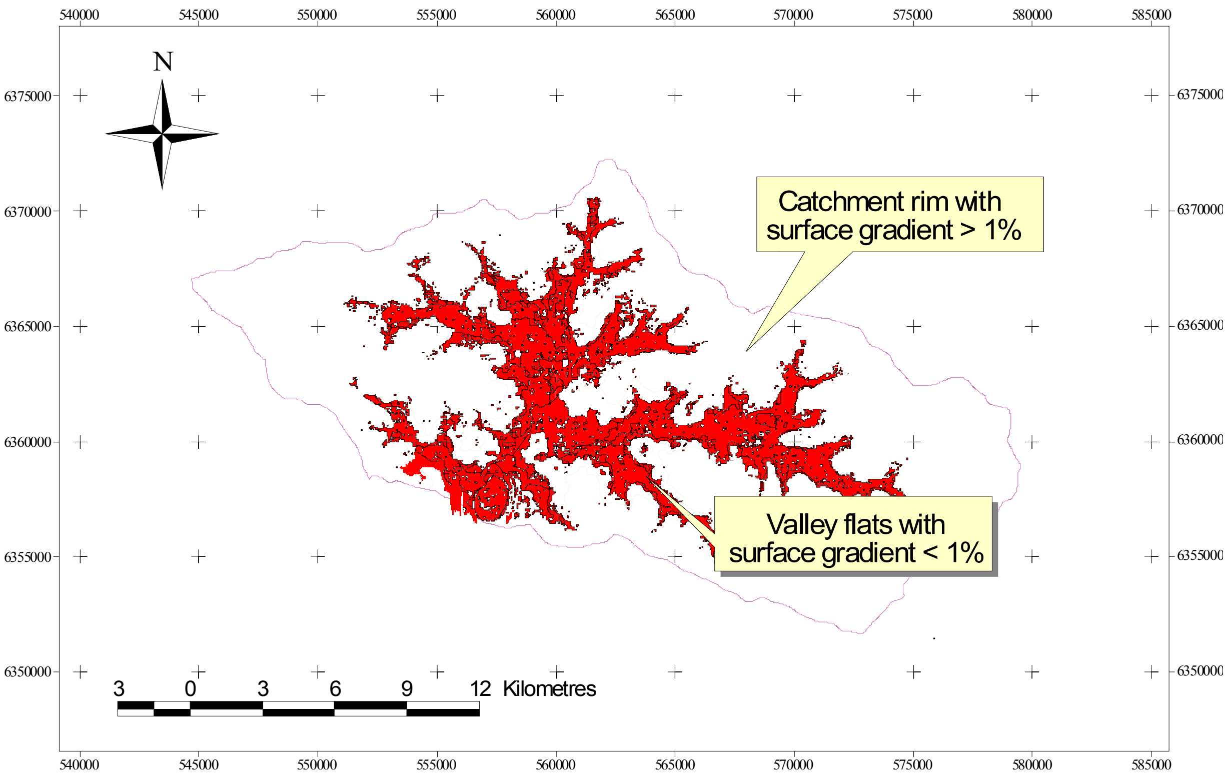

Figure 3. The valley flats extend over ~ 35% of the Toolibin catchment.

1.4 The impact of geology and weathering on groundwater flow

Rocks at the earth's surface are subject to physical disintegration or chemical decomposition. The weathering products can accumulate in places to form a soil, or can be transported by wind or water to accumulate as sediments, mainly in the low-lying area of the landscape. In addition, a primary mineral in igneous and metamorphic rocks subjected to weathering, may simply dissolve in situ, or a portion of it may re-precipitate to form secondary minerals. The later process tends to open pre-existing fractures and other pathways as well as create secondary pores within these rocks. The former is responsible for the initial deep weathering (40 to 60 metres) of saprolite within the catchment.

Weathering can take place both above and below the water table, but is generally slower in a saturated environment. The weathering processes in both saturated and unsaturated environments has an important control on the occurrence of groundwater. The differences in sub-surface mineralogy and the resulting variations in hydraulic parameters, such as hydraulic conductivity and porosity, have a profound effect on flow systems. Mineralogical heterogeneity affects the interaction between flow systems, the surface pattern of recharge and discharge areas, and the quantities of flow that are discharged through the catchment.

The distribution and the lateral extent of the rock types underlying the Toolibin catchment, such as dolerite dykes and granite rocks, influence groundwater occurrence and movement in the catchment. The water content of saturated materials derived from weathering of these rocks increases linearly with clay content, and are greater for the deeply weathered material derived from dolerite than that in granite rocks (McCrea et al., 1990). The materials derived from dolerite and granite rock are highly porous up to 50% porosity with mostly very small sized pores (less than

0.1μm). The high porosity of both rock types enables the weathered zone to accumulate a large volume of water. However, the predominantly small-sized pores, retard the transmission of water through these materials. Therefore, water movement is most likely to occur through macro-pores and preferred pathway channels that separate this impervious matrix (Johnston et al., 1983), the saprolite aquifer and palaeochannels (George, 1992).

The lateral contrast in porosity, hydraulic conductivity and basement morphology of the fresh and weathered material derived from dolerite dykes and granite rocks may have an important influence on local groundwater flow. Such lateral and vertical contrasts in hydraulic conductivity between the weathering products of dolerite dykes and the surrounding weathered granite rocks may promote or retard groundwater seepage to the surface. The orientation of dykes and fractures, relative position of sediments and depth of regolith all help to determine the nature and rate of seepage.

In addition, as the vertical hydraulic conductivity contrast is increased between the main aquifers (saprolite and palaeochannel) and the overlying confining or semi confining layers (lacustrine and finer textured sediments), the vertical gradients are increased and the horizontal gradient in the underlying aquifer is decreased. This translates to a relatively higher potential for upward leakage from the deeper aquifers, and therefore, the discharge area increases to compensate for the larger flow rates from the main aquifer. It may also relate to the location of discharge areas, where seepage at the break in slope may be head driven (pressure derived from the upslope discharge through saprolite) whereas valley floor discharge may be restricted, depending on the permeability and connectivity of overlying sediments. The area of discharge is then a function of supply (recharge) and evaporation (watertable depth).

2. Previous investigations

Toolibin Lake and its catchment is one of the most extensively investigated areas in terms of groundwater and surface water hydrology and salinity management strategies in Western Australia (NARWRC, 1978; Furness, 1978; Kevi, 1980, Martin, 1982, 1986, 1990; Stokes and Sheridan, 1985; Greenbase Consulting, 1991; GHD, Consulting, 1992; JDA Consultant Hydrologists, 1994; George and Bennett 1995; George 1998, Dogramaci, 1999; De Silva, 1999; Froend and Storey, 1996; Froend, et al., 1987; George and Dogramaci, 2000). The aim of most investigations was to understand the physical and chemical processes, which caused the deterioration of the catchment and lake environment (e.g. George 1998), or to test the usefulness of various salinity management options, to control salinity of the lake (Martin, 1982, 1990; George and Bennett, 1995).

The Committee for the Rehabilitation of the Northern Arthur River Wetlands (NARWRC, 1978) carried out the earliest investigation into the impact of the development of dry land salinity on the ecology of Toolibin Lake. The study concentrated on the impact of the saline inflow into the lake and highlighted the adverse affect of the regional and local watertable rise on the lake environment. It recommended that further investigations are conducted to better understand groundwater dynamics in the catchment and the establishment of a piezometer network to monitor water levels in the vicinity of Taarblin and Toolibin lakes.

Logs of shallow holes drilled at Taarblin Lake (three metres) and Toolibin Lake (2.5 metres) suggested that both lakes were underlain by clay containing saline groundwater (Furness, 1978). A seismic refraction survey indicated the presence of an unconsolidated layer less than three metres thick overlying indurated lake sediments or weathered granite. The interpretation of seismic data revealed that the depth to fresh granite bedrock across the lake ranged from 27 to 46 metres (Kevi, 1980).

Following these investigations, Martin (1982) tested the effectiveness of various management options including drain construction, groundwater pumping and tree planting to control watertable rise within Toolibin Lake. Due to the low hydraulic conductivity of the aquifers, the study concentrated on analysing the effect of tree planting along the western edge of the lake as a management strategy. The planting of native trees in 200-metre wide strips was recommended. The study suggested that this control measure might be the most suitable strategy to slow the watertable rise in the vicinity of the planted trees and within the lake.

In addition, a multi-port bore was installed in the lake to better understand the interaction between groundwater at various intervals beneath the lake and surface water (Martin, 1986). The study highlighted the importance of recharge contribution from the lake to the underlying shallow aquifer during wet seasons. The data showed that the lake had not reached a new hydraulic equilibrium and that the watertable would continue to rise, further threatening the survival of the vegetation.

Hydrological investigations into the impact of surface water inflow into Toolibin Lake have shown that the salinity of the water within the lake is caused by the input of salt from the salinised agricultural catchment (Stokes and Sheridan, 1985). This study concluded that the full salinity effect due to clearing of the catchment has not yet developed, and therefore, the salinity of the lake would continue to rise. A study by Greenbase Consulting (1991) into the salt balance of Toolibin Lake on the other hand suggested that the salinity in the lake might have reached equilibrium. However, the lack of long-term data precluded any confident determination of rising trend in salinity of the lake.

Based on later hydrogeological investigations, Martin (1990) recommended the construction of a well field comprising at least 25 bores to control watertable rise under the lake bed (300 hectares). These bores were each to have a discharge capacity of about 20 square metres per day (considered to be the maximum possible from the deeply-weathered saprolite) and the saline water discharged downstream to Taarblin Lake. The effect of additional saline groundwater discharge from Toolibin Lake to Taarblin Lake was deemed very small.

These studies provided essential background information to formulate the Toolibin Lake Recovery Plan (1994), which included implementation of various salinity management strategies across the Toolibin catchment and within the lake. The recovery plan recognised that the evaporation of surface water and eventual inflow of this highly saline water into Toolibin Lake, particularly in low flow periods, were detrimental to the health of the lake's vegetation. Therefore, the management of surface flows into the lake was considered a pre-emptive and integral part of the Toolibin Lake Recovery Plan (Toolibin Lake Recovery Team and Technical Advisory Group, 1994).

The Technical Advisory Group (TAG) completed the first stage of implementing urgent actions under the recovery plan with the design and construction of a large separator and diversion drain. The structures were designed to allow the saline low flows (less than six cubic metres per second - initial and end season flows) to be diverted, while allowing the fresher winter flows (more than 1,000 mg/L) to be directed into the lake. The construction of a channel to divert the saline inflow around Toolibin Lake was finished in 1995. In addition, and as part of the continuing work on surface water management, TAG commissioned JDA Consultant Hydrologists to investigate the effectiveness and design of an outflow control from Toolibin Lake. The construction of the outlet in 1999 allows for more frequent flushing of accumulated salt from the lake and results are presented later.

The Toolibin Lake Recovery Plan also acknowledged that the impact of watertable rise within the lake is at least as important as the influx of saline surface water in degrading the lake environment (Martin, 1990). However the early hydrogeological studies considered that the lake had developed on a thin bed of fine textured sediments (two to three metres) overlying 30 to 50 metres of deeply weathered Archaean bedrock (Martin 1986, 1990). Groundwater was noted within a moderately permeable saprolite aquifer (near the basement) and confined by the low permeability 'pallid' clays of the deeply weathered rocks (Figure 4). Therefore, an airborne magnetics survey (to locate dykes), additional piezometers and a new network of seven production bores were installed along the western side of the lake, the part most severely affected by salinity (George and Bennett, 1995). Although pumping only began in March 1997, results suggest that the water level dropped about 1.5 metres in the vicinity of the production bores.

A reconstruction of the hydrogeology of Toolibin Lake based on recent drilling of ~ 100 bores and an evaluation of SALTMAP and TEMPEST electromagnetics was undertaken (George and Bennett, 1995, George, 1998; Dogramaci, 1999, George and Dogramaci, 2000). Together with palynological evidence (De Silva, 1999), the data revealed that a deep and relatively transmissive palaeodrainage system exists in the lower catchment, extending at least five kilometres upstream of the lake. Above this point, only deeply weathered 'in situ' regolith and more recent shallow sediments can be found in the valley. This finding has had a significant impact on the design and the discharge capacity of the groundwater production bores within the lake. As a result, four new production bores were drilled, two into the palaeochannel sediments (estimated yield 300 kL per day) the two additional bores installed within the saprolite (80 and 14 kL per day).

The bores in the palaeochannel were installed to both induce vertical flow in the overlying lacustrine sediments and increase the drawdown, and hopefully effectiveness of older pumping bores located to the west. Results of the pre-pumping feasibility analysis and impact modelling (SKM 2000) indicated that over 80% of the lake floor would have water tables below three metres within 4,000 days of commencing pumping at full rates. Much of this impact would have developed within three years. Pumping at 70% of design rates commenced in summer 2001.

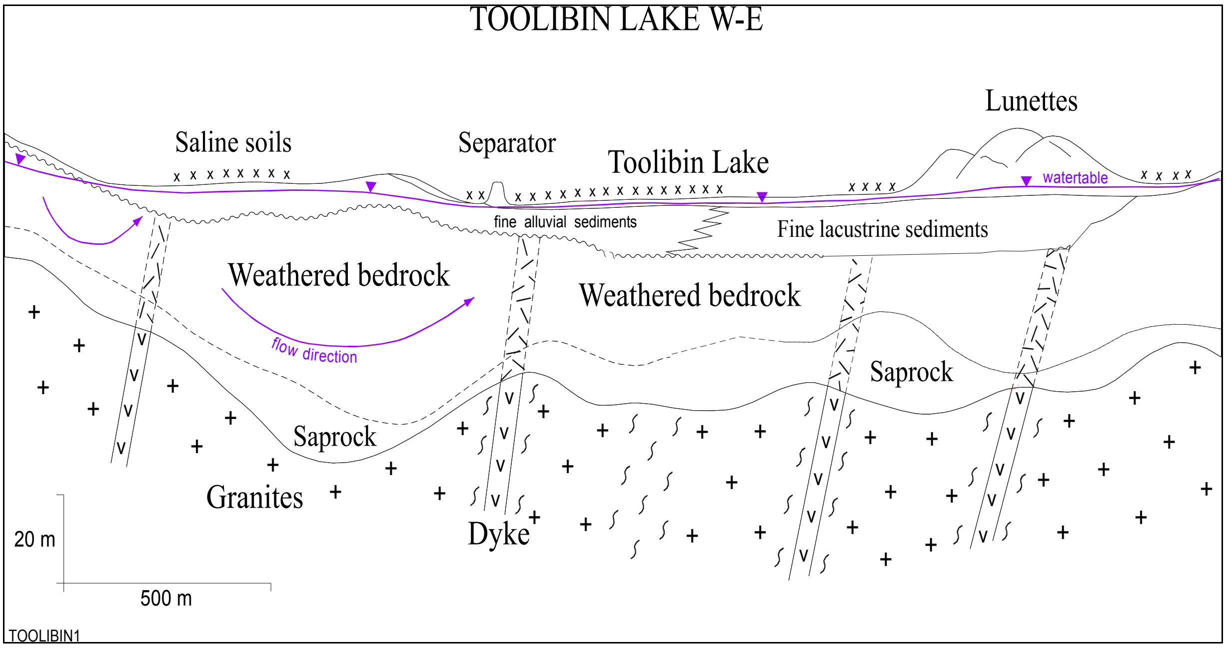

Figure 4. Cross section through Toolibin Lake based on the drilling program, which took place near the western shoreline and early conceptual model of Martin (1990). Note the sediments (more than five metres) and full weathered bedrock profile first assessed to be representative of all of the materials under Toolibin Lake.

3. Geological structures

In catchments of limited bedrock exposure, such as the Toolibin catchment, electromagnetic (AEM) and magnetic surveys may provide information on the lateral continuity and variability in the regolith. Magnetic images may reveal structures obscured beneath the regolith, while electro magnetics may contain information on their conductivity and depth. Such data was acquired as part of the National Airborne Geophysics Program (George 1998).

In general, the highly variable magnetic response of the Archaean basement rocks, and particularly a decrease in magnetic response in the south-eastern section of the catchment, may be attributed to the compositional change of the rock or magnetite destruction by the passage of fluids through nearby faults. There are several suites of Proterozoic dolerite dykes interpreted from the data (white lines on Figure 5). They are characterised by moderate to strong magnetic responses that probably correspond to a change in mineralogy (Figure 5). The dominant trends of these dykes are regional east-west trending, with a strong east-southeast and a dominant north-west phase corresponding to the regional mineral foliation (Chin, 1986). This foliation is responsible for curvilinear trends frequently observed in the dykes and is associated with metamorphism prior to their emplacement (Wilde and Walker, 1982). The north-west trending dolerite dykes are commonly thinner than their counterparts, and are particularly pervasive in the north-western section of the Toolibin catchment, where they cut and displace dykes with east west orientations.

Unlike the dolerite dyke's orientation, the regional faults (yellow lines on Figure 5) show north-south trends. These demagnetised zones represent major faults with intense magnetite destruction. Faults, along with dolerite dykes, locally control the position of modern natural drainage lines. However, frequently, the development of a superimposed drainage system simply reflects the contemporary land surface.

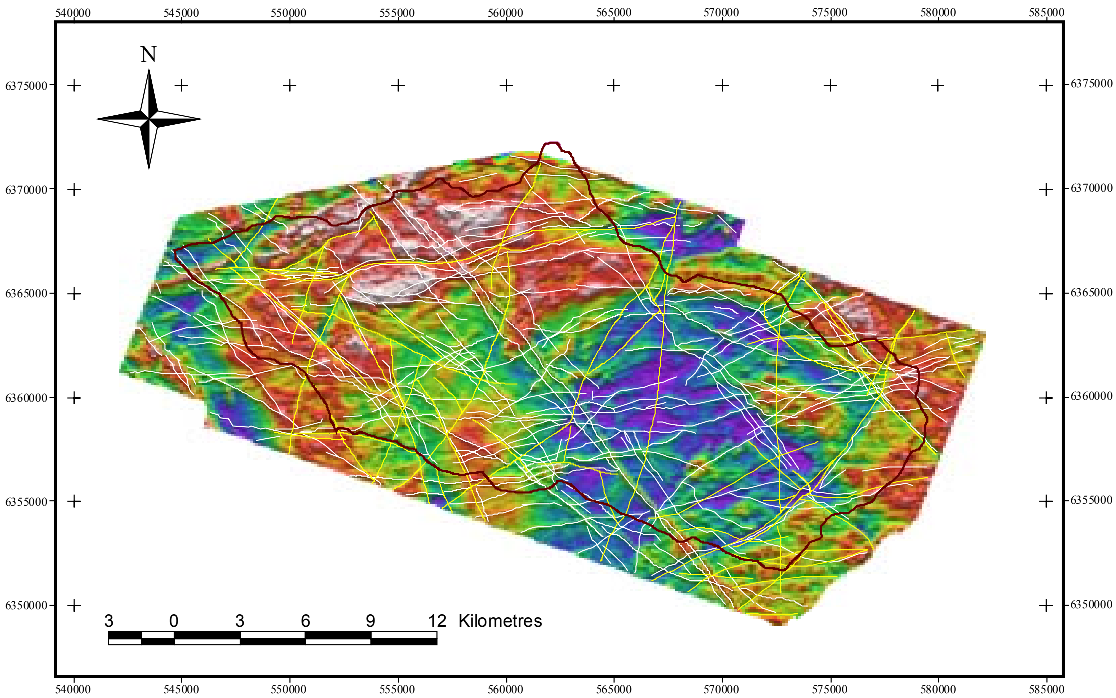

Figure 5. Aeromagnetic survey showing the regional faults (white lines) and dykes (yellow lines) within the Toolibin catchment.

The local trends of smaller dykes and faults generally mimic the regional alignments. Northwest-southeast trending dolerite dykes occur throughout the catchment and are displaced by north-east trending dykes. The disruption of the lateral continuity of many of the dolerite dykes may result from ‘pinching and swelling’ during intrusion, or magnetite destruction associated with geological events including emplacement, cooling, and the passage of alteration fluids during episodes of metamorphism and faulting.

The location of local dykes has been postulated to play an important role in current groundwater flow direction and the interrelations between intermediate and local flow systems within the catchment (McFarlane et al, 1989). However, based on the current location of salinity, and the location and orientation of the dykes in the catchment, they appear to be only having a limited impact on the development of groundwater seepage and salinity development. In the medium term, however, when more of the deeply weathered landscapes fill, dykes may play a greater role. Ultimately, however, when the catchment attains a new hydraulic equilibrium, such geological structures will have minimum impact on the watertable distribution or salinity development and groundwater budget (George et al, 2001, Dogramaci et al., 2003).

4. Catchment hydrogeology

4.1 Weathered profile

The weathering and erosion of the granites in the Toolibin catchment, over a long period of relative geological stability has resulted in several characteristic deep weathering and sedimentary profiles (George, 1992, George, 1998; DeSilva, 1999; Dogramaci, 1999).

The results from drilling about 100 shallow and deep bores across the catchment suggest that the deep weathered profiles in the Toolibin catchment can be described according to three main horizons (Figure 6 and 7). A lower horizon developed above the basement at the weathering front is characterised by the formation of saprock. The saprock develops in response to the fragmental disintegration of basement rocks rich in quartz and feldspar and is typically 'gritty' promoting the separation of individual mineral grains within the rock. Generally, saprock forms when up to 20% of the primary minerals weather to produce secondary clay minerals. The saprock in the eastern part of the catchment forms directly over unweathered basement because the basement rock contains a high percentage of mafic minerals, and/or a small average grain size of the primary minerals.

Above the lower horizon (saprock), primary mineralogy becomes less important as most primary minerals (other than quartz) weather to clay promoting the development of a sandy-clay horizon. This unit acts as a semi permeable layer in the catchment impeding the upward movement of deep groundwater from the saprock horizon to the surficial sediments that cover the majority of the valley flats. This in situ weathered profile is called the saprolite.

Surficial sediments of medium to coarse-grained quartz sands and clays, overlie the in situ weathered profile. Surficial deposits derived from alluvial, colluvial, and aeolian origin may have a higher hydraulic conductivity than saprolite (Quaternary alluvial phase). Although a veneer of surficial sediments may occur on the slopes at the rim of the catchment (more than three metres), they cover the entire valley flats on the main drainage line in the middle and lower parts of the Toolibin catchment.

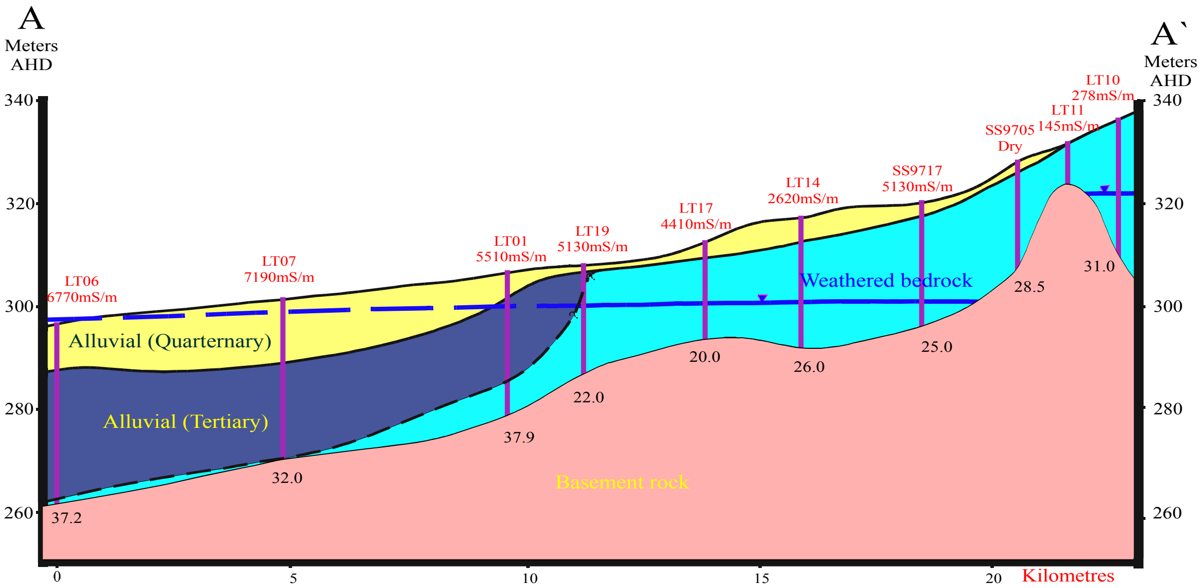

Figure 6. East-west AA` transect showing the hydrogeological setting of the Eastern section of Toolibin catchment. Refer to Figure 1 for the location of transects. The hydraulic gradient in the upper valley area of the catchment (eg between LT01 and SS9701) was near zero in 1997. This suggests highly localised recharge and an immature aquifer system.

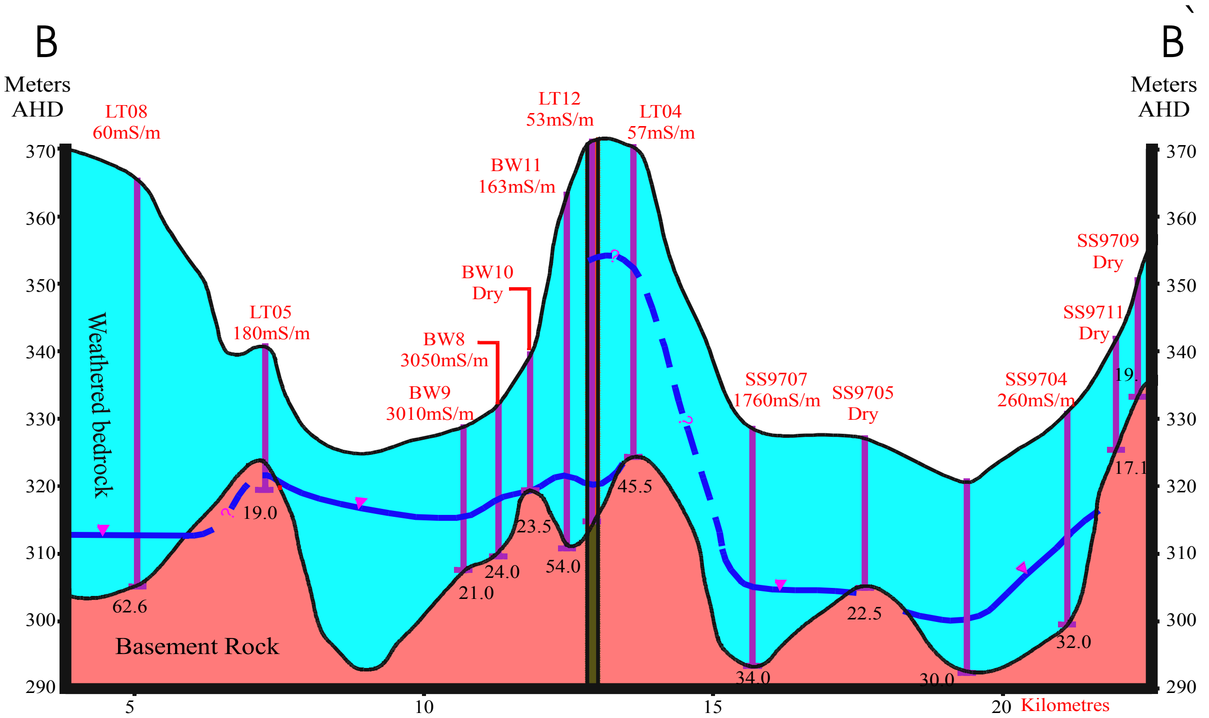

Figure 7. North south BB` transect showing the hydrogeological setting of the upper Toolibin catchment. Refer to Figure 1 for the location of transects. Groundwater levels and low salinity waters at LT04 indicates that a local recharge mound has developed in an area of sand plain.

The most important sedimentary system occurs at depth and is ascribed as the Tertiary Alluvial (palaeochannel) aquifer (Figure 6). Groundwater flow in this system occurs as a relatively highly transmissive palaeochannel system at the eastern margin of Toolibin Lake extending about five kilometre up slope north-east of the catchment (George and Dogramaci, 2000). The alluvial channel sediments mainly comprise sand and clay and are 20 to 30 metres thick. It is expected that the channel extends down the Northern Arthur River System. The sand particles are mainly sub-angular to sub-rounded quartz indicating a short travel distance and a rapid mode of deposition (Dogramaci, 1999). The alluvial sediments are overlain by up to eight metres thick lacustrine fine clay within Toolibin Lake.

4.2 Hydraulic conductivity and yield of the aquifers

The hydraulic conductivity of the material comprising the main aquifers in the Toolibin catchment varies as a consequence of the mineralogy and location in the landscape. The slug test analysis suggests that there are 30-fold differences in hydraulic conductivity between the top and the base of the surficial sediments. The hydraulic conductivities in the catchment decrease with depth, particularly in the middle of the valley flats (McFarlane et al., 1989). In these areas, the groundwater flows are likely to be slow in the basal units given the low gradients and low hydraulic conductivity (McFarlane et al., 1989).

Conversely, the hydraulic conductivity of the weathered profile up slope of the valley flats at the catchment rim increases with depth, suggesting that the saprock horizon has a relatively higher hydraulic conductivity than the overlying saprolite (George, 1992; Clarke et al., 2000). Hydraulic conductivity values are higher for the palaeochannel sediments than the adjoining deeply-weathered granite and range from 0.8 to two metres per day and 0.02 to 0.06 metres per day respectively (Dogramaci, 1999).

Pumping tests for bores in the palaeochannel and the weathered granite aquifer indicate that yields from the weathered granite aquifer are highly variable ranging from more than 10 m^3 per day to 150 m^3 per day (George and Bennett, 1995, De Silva, 1999; George, 1998; Dogramaci, 1999). This contrasts with consistently higher yields from palaeochannel sediments where yields range from 200 m^3 per day to 350 m^3 per day (Dogramaci, 1999).

4.3 Water table and groundwater head in Toolibin catchment

The groundwater head is the pressure in the deep groundwater (located in the partially-weathered granite just above bedrock or in the palaeochannel sediments) expressed as height above or below ground and it is measured as a water level in deep piezometers. By contrast, the 'watertable' is the surface on which the water pressure in pores of the aquifer is exactly atmospheric. The level at which water stands in a shallow bore (less than three to six metres at Toolibin) reveals the location of this surface.

When the groundwater head is higher than the watertable, there is a tendency for an upward flow from the deep groundwater toward the surface. Conversely, when the watertable is higher than the groundwater head, there is a potential for recharge to the deep groundwater. The potential for interaction of shallow and deep groundwater therefore can be determined from measurements in an adjacent monitoring bore and piezometer.

In the hill slope saprolite and valley floor quaternary sediments, there is relatively little difference between heads in the deeper aquifers and watertable levels. This indicates surface and deeper aquifers are relatively well connected throughout most of the Toolibin catchment. However, the head of the deep aquifer on the eastern side of Toolibin Lake (overlying the palaeochannel) is within 0.5 metres from the lake bed while the water table in the shallow aquifer is about two to three metres below the lake bed. Previous studies have indicated that head and water table depth are dependent on the water level within the lake (Martin, 1990). The interaction of groundwater and water within the lake is discussed in the following section.

4.4 Groundwater occurrence in Toolibin Lake

The early hydrogeological conceptual model for Toolibin Lake consisted of a thin layer of clay (two to three metres thick) overlying the regional groundwater within the weathered granite (Furness, 1978). The hypothesis to explain the occurrence of fresh water within the lake, was based on the assumption that the clay layer was acting as a barrier to the deep groundwater flow and preventing upward leakage from the highly saline groundwater into the lake (Furness, 1978).

The later re-interpretation of the groundwater and surface water levels within the lake (Martin, 1986, 1990) highlighted the dynamic interrelationship between groundwater and surface water. The analysis showed that Toolibin Lake acts as both recharge and discharge zone depending on the depth of water in the lake. The lake acts as a recharge zone during period of prolonged inundation when the water level in the lake is higher than heads in the deep aquifer. Conversely, when the lake is dry there is potential for discharge from deeper groundwaters to the soil surface and vegetation's root zone. The actual rate of discharge is dependent on the head and aquitard permeability. In the western margins of the lake, sediments are both thinner and more permeable than those to the east. Sandy alluvium occurs at depths greater than three metres and is overlain by highly sodic medium textured clays, and sandy clays (George and Bennett, 1995). Discharge occurs from the alluvium and deeply weathered zone aquifers, and the watertable and piezometric head are at or near the soil surface (Figure 8).

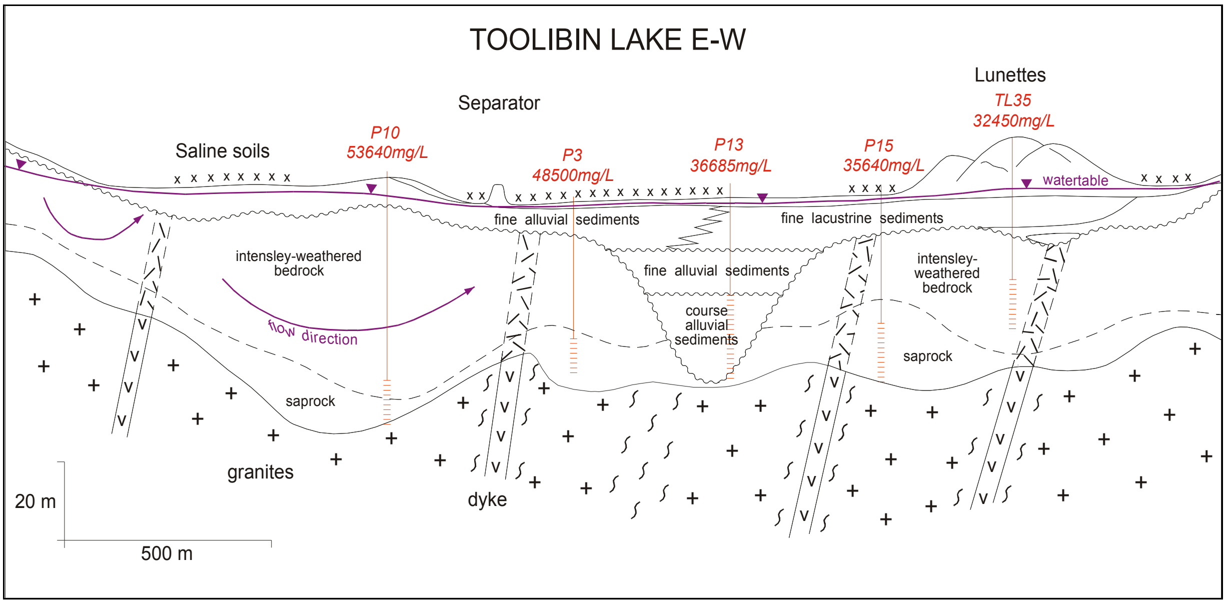

Figure 8.West to east Toolibin Lake cross section showing the hydrogeology of the palaeochannel within the lake (George and Dogramaci, 2000) and interactions with superficial and deeply weathered materials. The position of dykes is interpolated and they are not to scale.

The existence of a deep and relatively transmissive palaeodrainage system in the middle of the lake also has a major impact on the variation of the deep groundwater head. In the eastern part of the lake, a layer of fine lacustrine clay approximately eight metres thick, confines groundwater in the palaeochannel. Here the deep groundwater head is higher than the shallow watertable suggesting potential for upward leakage. Lacustrine sediments further east of the lake, underlying the lunettes, act as a semiconfined layer due to a higher sand content, resulting in lower head difference between the shallow watertable and the deep groundwater within the weathered profile. Saline seepage at the interface between the sediments and the lake floor appears to be principally from a perched aquifer linked to the lunettes (between one and three metres below the lake floor), although some deeper aquifer contribution may also be occurring.

The west-east hydrogeological section (Figure 8) of Toolibin Lake depicts the principal water-bearing formations in the lake and surrounding area. Groundwater within the palaeochannel, flows longitudinally towards Toolibin Lake from the Toolibin Flats (McFarlane et al, 1989). The position of the channel was marked by interpretation of the electromagnetics (SALTMAP and TEMPEST data) and showed a characteristic AEM signature (George 1998). The palaeochannel enters the lake 200 metres east of P1, turning sharply between a bifurcated dolerite dykes. Drilling at P1 and P1b indicated a steepening bedrock profile but no sediments. Pumping bores P11 and P13 encountered the deep channel sands, whereas at P12, in an area of poor AEM differentiation, but within the channel AEM pattern, lacustrine materials overly variable textured and lower permeability sediments. The electromagnetics suggests that the palaeochannel has a lithology and permeability that may vary along its length. Alternatively, drilling at P12 and nearby, missed the channel.

Aquifer tests and modelling (SKM, 2000) suggests that the palaeochannel is bounded by, and connected to, the weathered granite aquifer. The rate of groundwater flow between the two aquifers is primarily controlled by the hydraulic conductivity of the weathered granite and the difference in head between the two aquifers. The shallow groundwater occurs in the surficial sediments throughout the catchment and in the lacustrine sediments within the lake.

5. Groundwater flow

The accuracy of water level gradients across the catchment depends on the number of bores in the catchment and the surveying accuracy. In locations where the data are scarce or unavailable, the temporal watertable configuration may be estimated by extrapolation of the water level in any given point to the surrounding areas. The heterogeneity of the hydraulic properties of the aquifers results in the development of complex groundwater flow systems (i.e. intermediate and local flow systems), especially near the palaeochannel. While, extrapolation of the water level may not represent the true water level across the entire catchment, it does provide information on the likely groundwater flow direction and the probable discharge areas in the catchment.

The measured groundwater heads and watertable are highly variable, ranging from positive head above the surface to 25 metres below the surface. The records of 50 monitoring bores (surveyed to an accuracy of ~ 0.025 metres, George, 1998) and with recent water level information were used to construct a contour map of the shallow watertable in the catchment. The groundwater contour map reveals that while the watertable under the valley flats generally mimics the topography, in the eastern catchment (Scriveners Soak), valley groundwater gradients are near zero. This reflects localised recharge, poor lateral connectivity and the dominance of vertical fluxes. In the uplands the water level varies according to local hydrogeology. The watertable in the south-eastern part of the catchment in sand plains is 16 to 25 metres below the surface. In contrast, the data from bores drilled in the north-western part of the catchment suggest that groundwater is close to the surface and does not occur at sites with shallow basement rocks.

Based on the generalised watertable contour map (Figure 9) the groundwater flow is from the catchment rim toward the valley flats eventually discharging into Toolibin Lake. Lateral flow rate calculated by Darcy's law varies from 0.01 to one metre per year, depending on the groundwater gradient and hydraulic conductivity. Based on these numbers, the travel time for regional groundwater from the catchment rim to Toolibin Lake may be from 100,000 to millions of years. Actual travel paths are dependent on recharge rates. When recharge exceeds the effective aquifer transmissivity, discharge will take place near recharge areas. George et al (2001) determined that the critical recharge value at Toolibin was one millimetre per year. At values of less than that, discharge areas would develop near the flats or Toolibin Lake. At higher values, large areas of the catchment would be required to discharge groundwater (see Modelling Section).

Carbon dating of groundwater under Toolibin Lake indicates ages less than 10,000 years. This may indicate that the groundwater under Toolibin Lake within the palaeochannel and weathered profile is recharged within five to 10 kilometres up gradient of the lake (the area where the AEM suggests the palaeochannel extends), and/or that the ground water represents a mixture of palaeo and recently recharged water. The later is most likely.

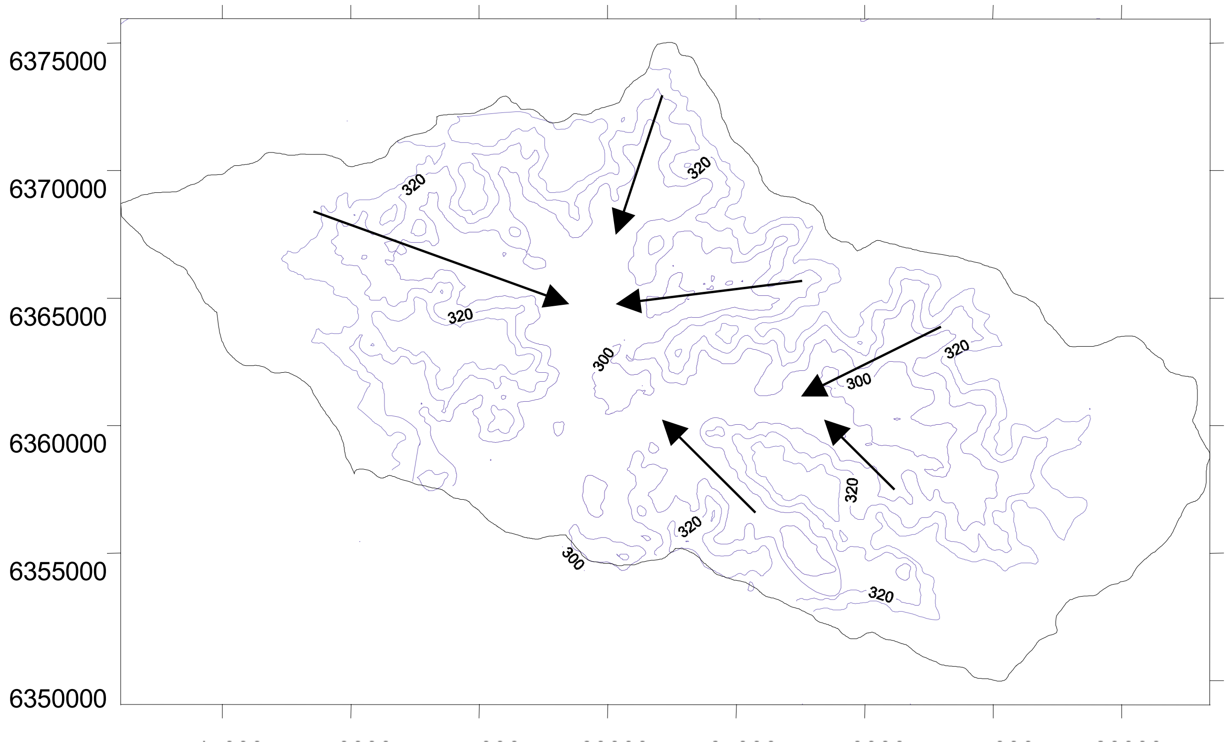

Figure 9. Groundwater contour map showing elevation of the watertable across the catchment. Gradients from Scriveners Soak to the Flats and Toolibin Lake are less than one metre in 20 kilometres.

Although water level contours are important to delineate lateral groundwater flow direction, they do not show transient or long-term patterns in recharge and discharge. The temporal variation of the water level can provide information on the state of catchment equilibrium and how the volume and flow rates of the deep groundwater are responding to changes in hydrological cycle and land use.

5.1 Water level trends

The groundwater level may fluctuate in response to short-term variations in rainfall or annual changes in storage. Water level trend analysis using 'Hydrograph Analysis: Rainfall and Time Trend' (HARTT) model (Ferdowsian et al., 2000) can be used to differentiate between the effect of rainfall fluctuations and the underlying trend of groundwater level over time.

Rainfall is represented as an accumulation of deviations from average rainfall. Based on monthly rainfall and observed groundwater level, the HARTT model predicts the trend of the groundwater level.

Fourteen piezometer records representing groundwaters in the catchment rim, valley flats and lacustrine sediments within Toolibin Lake were selected for water level trend analysis (Figure 10). The piezometers representing water level in the saprock aquifer (SS9701D, SS9704D and SS9706D) are all located in the eastern part of the catchment. The rest of the bores are spread across the catchment including four piezometers in Toolibin Lake (TL16, TL 19, TL 22, and Tl 26). The average rainfall record for 23 years covering the period from 1977 to 2000 was used in the model.

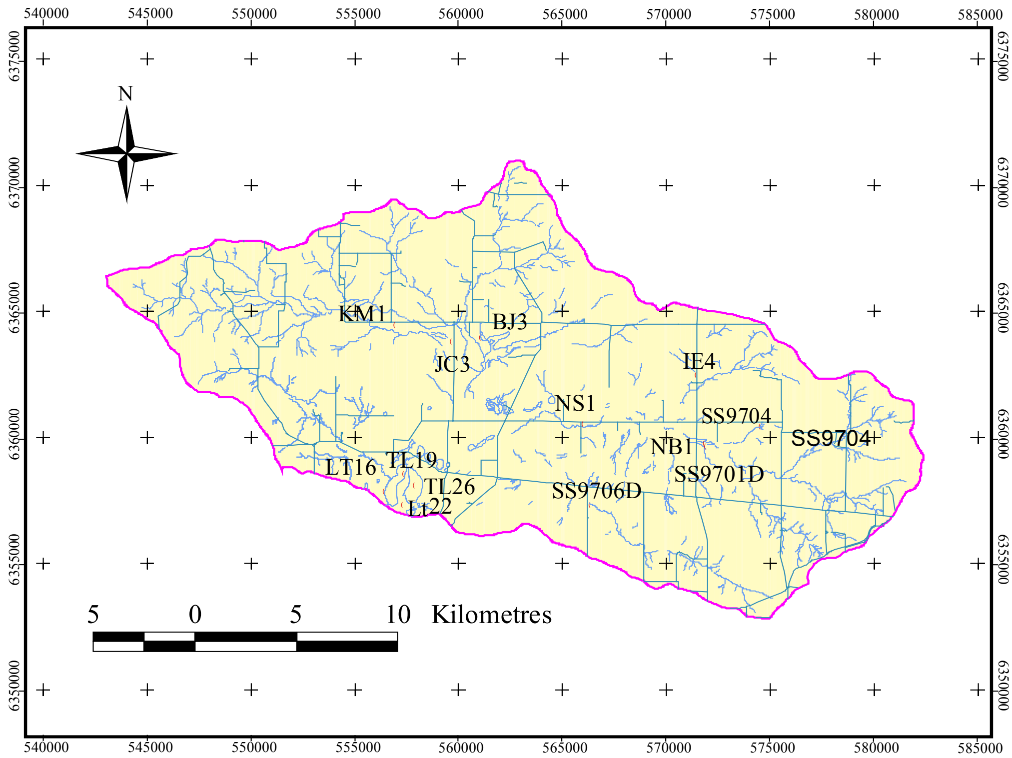

Figure 10. Monitoring bore locations chosen for HARTT analysis superimposed on the cadastral map of Toolibin catchment.

The shallow bores (Figure 1 to 7 – Appendix 1), located in the valley flats, have a relatively low rate of rise. The low correlation coefficient ($R^2 = 0.8$ to 0.65) calculated for these bores can be explained by average monthly rainfall residual and time.

Figures 1 and 2 (Appendix 1) are representative of shallow bores that are located in the middle of the valley flats and close to discharge sites. These bores have a slight falling trend with R^2 of 0.65, and reflect short-term inter-seasonal variability. The best fitting lag (the response of water table to rainfall) for both bores is zero, suggesting an immediate response to rainfall. The correlation coefficients for most of the shallow bores (more than three metres deep) are low, indicating that processes not taken into account by hydrograph analysis such as the direct evaporation of groundwater, affect groundwater level fluctuations.

The water level trend analysis for the three deep bores (Figure 8, 9 and 10 – Appendix 1) in the eastern part of the catchment within sand plains, shows a rapidly rising watertable. The pattern of rise is well captured by the estimated regression model ($R^2 = 0.96$). The best fitting lag length for these bores are 27, 28 and 15 months respectively. This does not equate with the time taken for the rainfall to enter the aquifer as a recharge pulse. This may only take a few months.

There is a clear trend for increasing lag with increasing depth, with between-bore variation also increasing with depth. The estimated lag lengths shown are the lags that provide the best statistical correlation, and represent the maximum impact. The delay in this maximum impact is primarily dependent on hydraulic parameters of the aquifer and groundwater gradient. The variation in groundwater lateral flow from the recharge area to the bore site is captured as a lag time revealed in the bores.

Rate of groundwater rise for deep bores tends to be substantially greater than for shallow bores (Appendix 1). This partly reflects the greater role of discharge in landscapes with shallow bores; this discharge is effective in removing water from the system and slowing the rate of groundwater rise.

Furthermore, the water level trend analysis indicates that the HARTT model better predicts the trend in deeper bores than in the shallow bores. The four main observations with important implications for the long-term water quality and area of dry land salinity within the Toolibin catchment can be drawn from water level trend analysis:

1. The vertical flow is dominating the groundwater movement. Water levels are rising in the catchment rim by approximately 0.5 metres per year, which is similar to, or relatively faster than the lateral groundwater flow (0.1 metre per year to one metre per year). However, because the maximum depth of the water level in the catchment is ~ 25 metres below the surface, it takes approximately 50 years for the groundwater to reach the surface. Laterally flowing groundwater on the other hand, takes thousands of years to travel from the top to the bottom of the catchment.

2. The dominance of the vertical flow in discharge areas will cause an accumulation of salt in the soil profile and alter the salt balance of the catchment. Although upward trends are evident in some of the bores the average rise in the shallow bores within the valley flats is zero. Additional salt loads will come from an expanding area of dry land salinity.

3. The groundwater pumping has had impact on controlling the watertable within Toolibin Lake. Draw downs are highly variable, but range from greater than five metres near pumping bores to more than two metres across large areas of the western lake area. Piezometric water level reductions of greater than two metres have also occurred under the palaeochannel and some evidence exists of an accelerated rate of drawdown in western bores. Dry conditions over the past two years make HARTT analysis problematic.

4. The water level trend analysis suggests that the full extent of land salinisation has not been developed in the catchment particularly within the valley flats. Discharge areas can be expected to expand up gradient encompassing the minor tributaries adjoining the Toolibin Flats, eventually including most valley floor soils, leading to higher discharge rates and increased salinity of surface water.

Catchment salt balance and surface water trend analysis can provide further information on trends in surface water salinity and increase confidence in the predictions of salinity trends for surface water discharging from the catchment. Modelling is used later to provide evidence of the timing and magnitude of the effect of water level increases.

6. Hydrology of Toolibin Lake

Toolibin Lake is ephemeral, only filling in years of above average rainfall. When full, the lake is about three-square kilometres in area, and two to three metres deep. The lake volume and surface area relationship derived from the calculation of a surface area-height relationship from contour map of the lake is shown in Figure 11.

The lake bed is relatively flat with many shallow, 0.2 to 0.5 metres depressions and mounds in the centre. These micro-topographic features influence the water movement when the volume of the lake water is low. The accuracy of volume area relationship diminishes at low water volumes.

Surface water inflow to Toolibin Lake is from the catchment of the Northern Arthur River (435 square kilometres) to the north-northeast of the lake (Gauging station No 609010) and from a smaller north-west creek catchment (40 kilometres), draining directly into the north west corner of the lake. Lake out flow is through an overflow channel to the south via a series of smaller lakes to Taarblin Lake. These conditions were changed following construction of the separator (1994) as Northwest Creek water was directed into the separator, bypassing the lake.

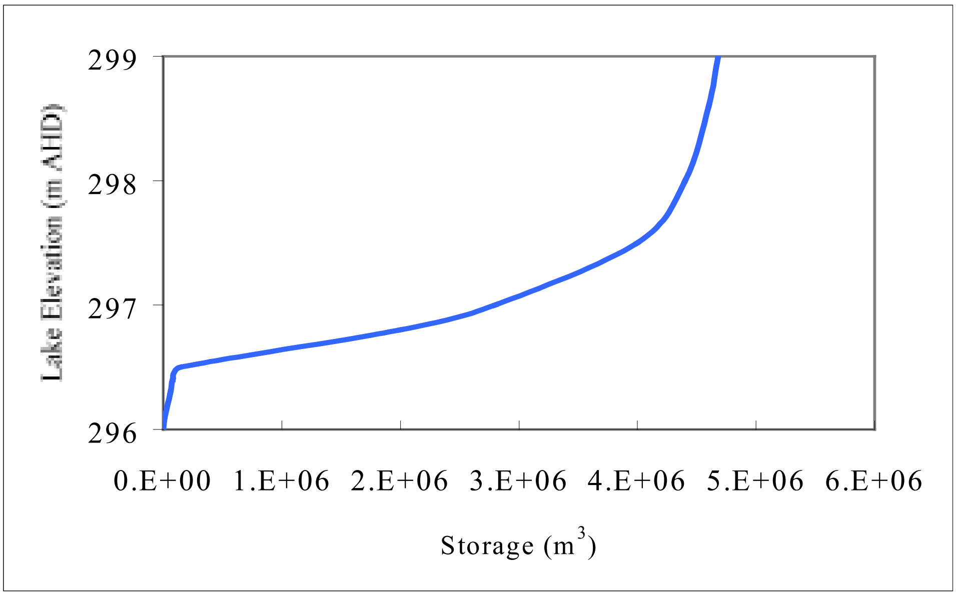

Figure 11. Relationship between elevation of water within the lake and storage volume in m^3.

Records of the flow from Northwest creek are available only for the three years (1982 to 1984). There is a poor correlation between the inflow from North Arthur River and Northwest Creek due to the short inflow record and variation of the rainfall intensity in different parts of the catchment. Nevertheless, an estimate can be obtained from the relative catchment area. In the absence of longer flow records, it is assumed that in most large inflows the Northwest creek contributes ~ 10% of the flow as the proportion of the catchment area. Northwest creek is approximately 10% of the Toolibin catchment.

The stream flow into Toolibin Lake is highly variable (Figure 12) ranging from periods of no flow for four years, to annual runoff ten times mean annual flow. The annual rainfall to runoff relationships (Figure 13) shows that annual rainfall of approximately 350 millimetres is required before significant stream flow is observed.

By assuming that the change in soil water storage in the catchment is negligible (from year to year) then annual catchment evapotranspiration can be assumed to be equivalent to the annual catchment rainfall, minus annual stream flow. The catchment evapotranspiration to rainfall relationship shows that significant departure from the one-to-one line doesn't occur until over 450 millimetres of annual rainfall (Figure 14). The potential evaporation or evapotranspiration would be much higher than the actual evapotranspiration, which in most years is equivalent to the rainfall.

The lake volume and salinity records suggest that the salinity of the lake ranges from 1,800 mg/L for the volume of $2x10^6$ m^3 to more than 10,000 mg/L for the volume less than $0.5x10^6$ m^3. The volume of water decreases with steepening recession into summer, suggesting an increasing evaporation influence. The calculation of lake salt load and average salinity of Toolibin Lake will be influenced by variation of inflow with time. The salinity of the inflow varies from less than 240 mg/L to 26,000 mg/L (Figure 15) and depends mainly on the time and volume of the flow.

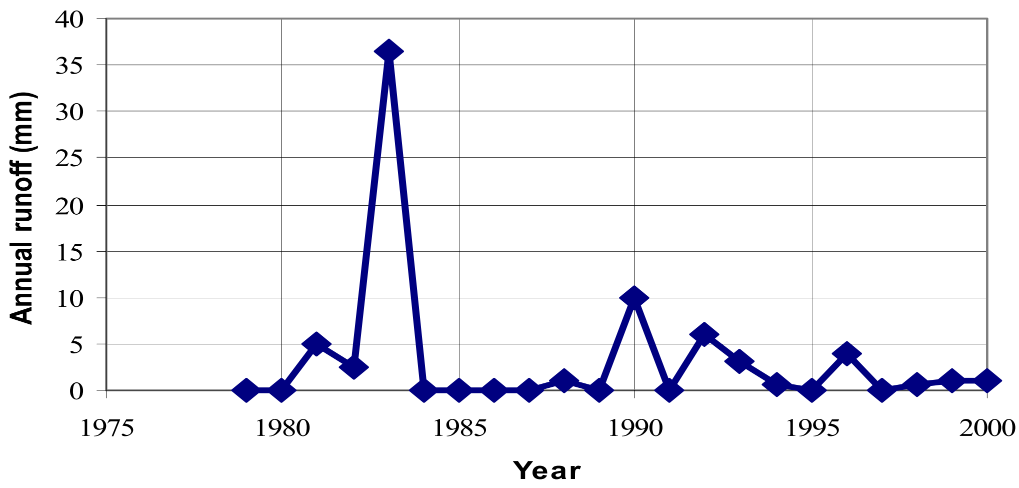

Figure 12. Annual runoff with time for Northern Arthur River (609 010).

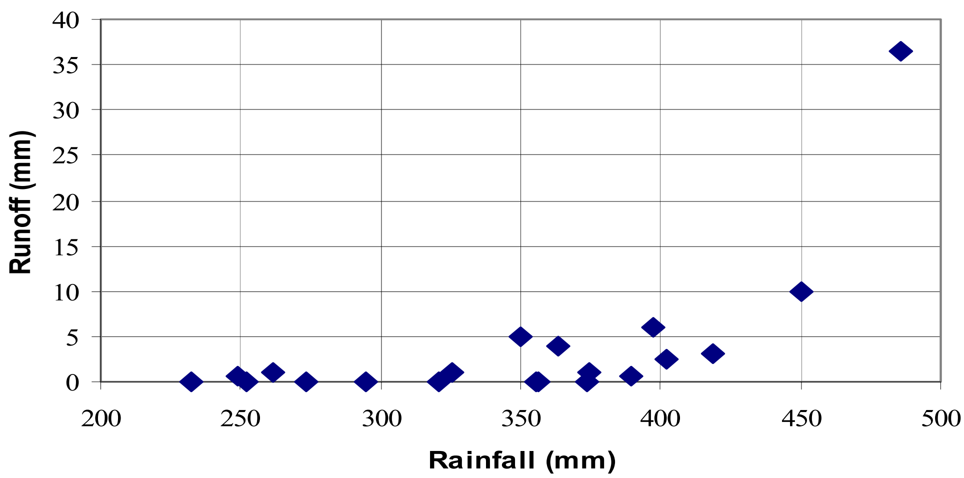

Figure 13. Annual runoff to rainfall relationship for Northern Arthur River (609 010).

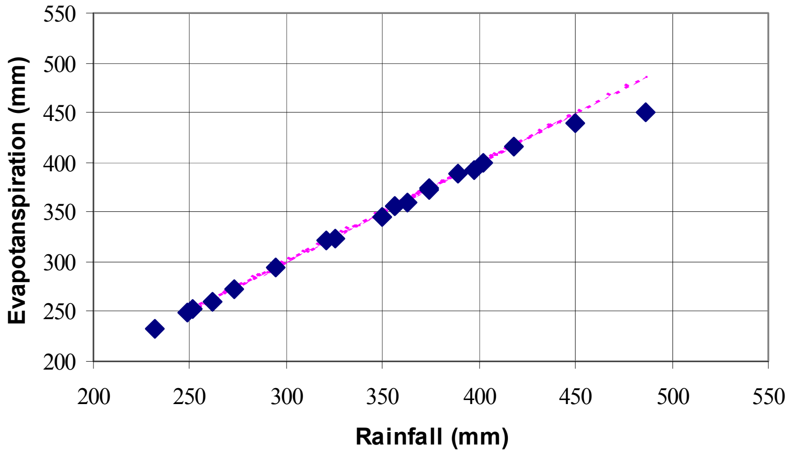

Figure 14. Annual evapotanspiration to rainfall relationship for Northern Arthur River (609 010).

Statistic	Flow (ML)	Salt Load (t)	Salinity (mg/L TDS)
10 Percentile	0.03	0.05	400
Median	272	489	2110
90 Percentile	2490	2980	4200
Mean	1405	1290	3500

Table 1. Flow and salinity summary for Toolibin Lake Inflow (609 010).

The summary of the flow and salinity of the inflow (609010) is shown in Table 1. The analysis of the data suggests that the median flow is 272×10^3 m^3 and the salinity is 2,110 mg/L. Only 10% of the flow is higher than $2{,}490 \times 10^3$. The difference between median and mean values is mainly due to the large variation in the flow rate and salinity of the inflow to Toolibin Lake.

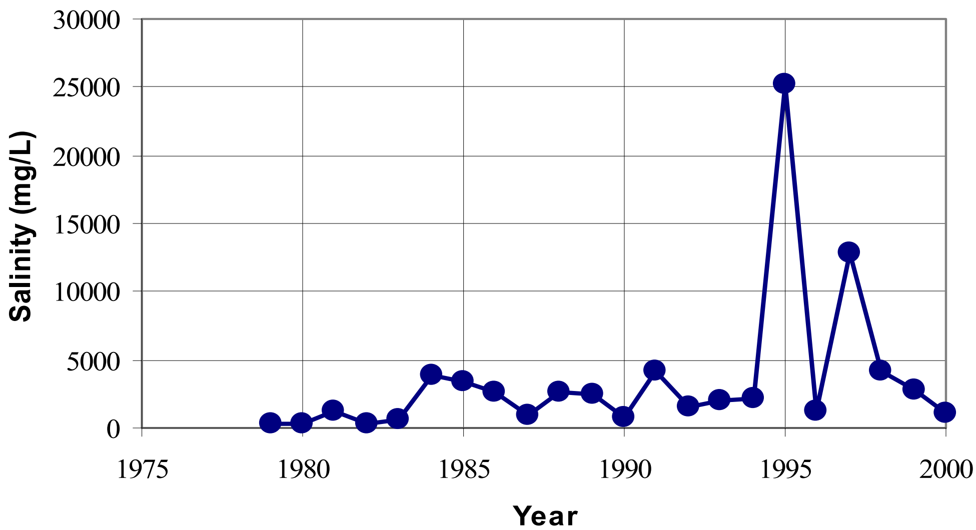

Figure 15. The annual average salinity record of the inflow (609010) to Toolibin Lake.

The surface water flow volume and salinity highlights the significant contribution of low volume saline water inflow to the salt balance of the lake. In 1994 and as part of the Toolibin Lake Recovery Plan a separator was installed so the saline water flow to the lake can be controlled and diverted. Since the installation of the separator approximately 3,500 to 4,000 tons of salt has been diverted from the Toolibin Lake.

An outflow regulation system has also been installed to allow lake water to be discharged into the separator channel and to allow management of water depth (vegetation survival and recruitment). This allows some control when the period of inundation is too long or additional lake bed leaching is required. Recent surface water modelling results (J. Davies, pers. comm.) suggest that the diversion channel and single outlet pipe will result in 80% reduction in the increase of salt load in the lake. About 60% of this reduction is due to the diversion channel and 20% due to the outlet pipe. The 80% reduction in salt load equates to minimum 8,000 tonne reduction over the 19-year modelling period (1979 to 1997).

6.1 Catchment salt balance and salinity trend analysis

The salt output/input (O/I) ratio from a catchment can be used as an indicator for land and surface water salinisation (Peck and Hurle, 1973). Generally, the O/I ratio for a catchment such as the Toolibin catchment would be expected to be near one prior to clearing of native vegetation. The unity of O/I value suggests salt equilibrium. Typically, however, salt is being accumulated in the soil profiles of native vegetation in this rainfall zone and O/I ratios are less than one. Clearing may initially mobilise salt stored in the soil profile through increased recharge causing additional leaching of salt into the groundwater. The elevation of the hydraulic head in the deep aquifers, with the subsequent increase in the groundwater results in an increased discharge volume together with its salt load. The increased saline groundwater seepage into the surface water increases the rate of output of salt relative to salt input via rainfall.

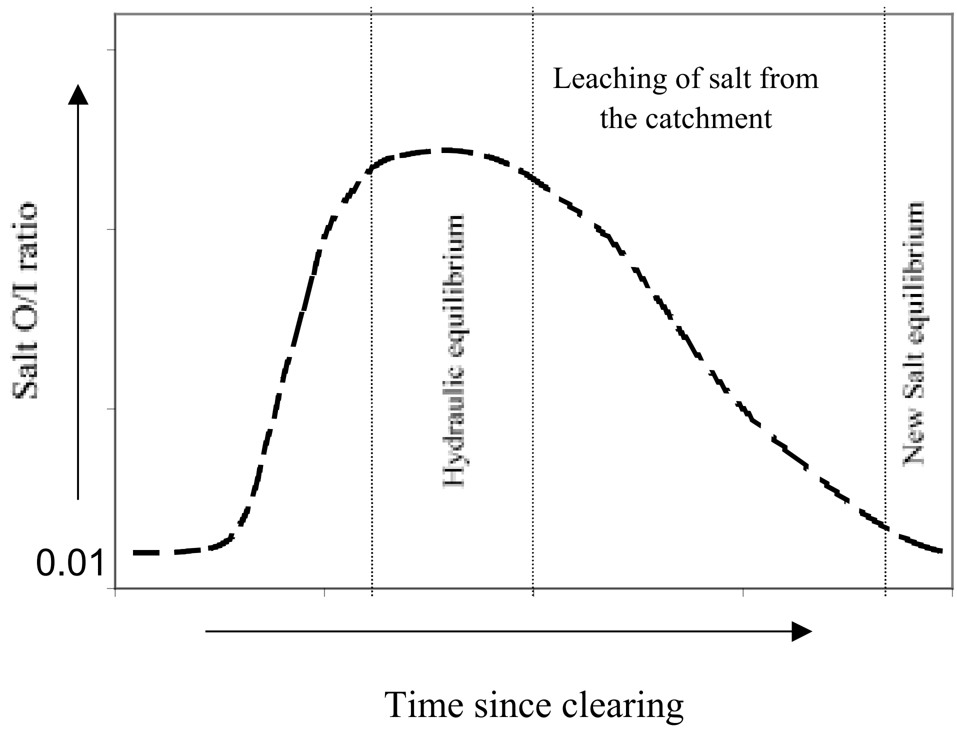

Figure 16. Diagrammatic relationship between salt O/I ratio and hydraulic and salt equilibrium.

The new hydraulic equilibrium (recharge = discharge) will precede the achievement of salt equilibrium, with the lag time depending on the leaching rate and the magnitude of the stored salt (Figure 16). This is due to additional water entering the system long before surface discharge of salt and water begins. Assuming all of the salt stored in the soil profile is mobile, a new salt equilibrium will be achieved when the stored salt is fully leached from the catchment. At the new hydraulic equilibrium, the salt flux discharging from the catchment will be more due to the significant increase in water discharge with its accompanying salt flux. The highest salt O/I ratio would be expected at the time of the new hydraulic equilibrium.

The estimated time required to restore salt equilibrium in the south-west of Western Australia ranges from 30 to 400 years for the catchments with the mean rainfall range of 1,120 to 490 millimetres per year respectively (Peck and Hurle, 1973). An estimated time for hydraulic equilibrium in a fully cleared catchment with mean rainfall of 1,030 millimetres per year was 64 ± 50 years and that 87% of the excess salt would be removed from the catchment in about 90 years (Macpherson and Peck, 1987). It is also expected that for catchments with mean rainfall less than 500 millimetres, the time for salt equilibrium will be in the order of thousands to tens of thousands of years, depending on rainfall, hydrogeological setting, salt store, catchment size and amount of vegetation clearing.

Most of the sub-catchments comprising the Blackwood River Basin, especially in western areas with greater relief and rainfall, are characterised by higher salt output via surface water relative to the salt input via rainfall. The salt O/I ratio calculated for 25 sub-catchments in the Blackwood River Basin ranges from eight to 30 (Bowman and Ruprecht, 1999). Conversely, the record of O/I ratio for the inflow to Toolibin Lake ranges from 0.0001 to seven depending mainly on the volume of runoff (Figure 17). For example in 1983 where the highest runoff was recorded for inflow, the O/I ratio is the highest since records commenced in 1977. Unlike the majority of catchments in the lower Blackwood River Basin, the yearly average O/I ratio for the Toolibin catchment is close to unity (Bowman and Ruprecht, 1999). This is a very peculiar characteristic given the average rainfall, high salt storage and the high percent of clearing (about 95%) in the catchment. The O/I =1 suggests that the response of the catchment to the early clearing has not been fully expressed in terms of land salinisation.

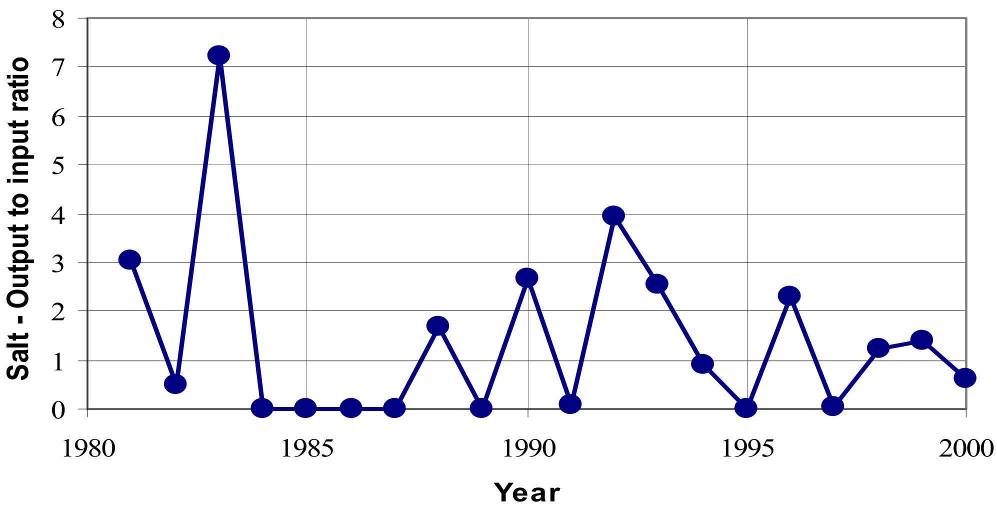

Figure 17. Variation in salt output to input ratio. Note that the highest recorded O/I corresponds to 1983 high inflow (Figure 12). In several years the ratio was effectively zero.

6.2 Surface water salinity trend

The measured daily inflows into Toolibin Lake from Arthur River gauging station 609109 for the years 1980 to 2000 are highly variable, ranging from ~ 0.001 millimetres per year to 70 millimetres per year (over five orders of magnitude - Figure 18). The rainfall record obtained from Wickepin meteorological station located 20 kilometres north of the catchment varies from the minimum of 238 millimetres per year to the maximum of 485 millimetres per year (Figure 13). The stream flow and mean flow-weighted salinity records for the Arthur River inflow are summarised graphically as annual values in the Figure 18. The data falls into two different groups. There is a strong inverse relationship between flow-weighted salinity for annual flows more than one millimetre but scattered data for the annual flows less than one millimetre.

Because of the inconsistency in the relationship between the rainfall and flow weighted mean salinity, and the large variation in both rainfall and flows, the surface water salinity trend has been determined each year. By fitting the equation of a power curve to the flow and salinity values from the four years before and the four years after the year in question, nine data points are derived. The resultant graph of salinity at mean annual flow reveals trends in time that are largely independent of variations in rainfall.

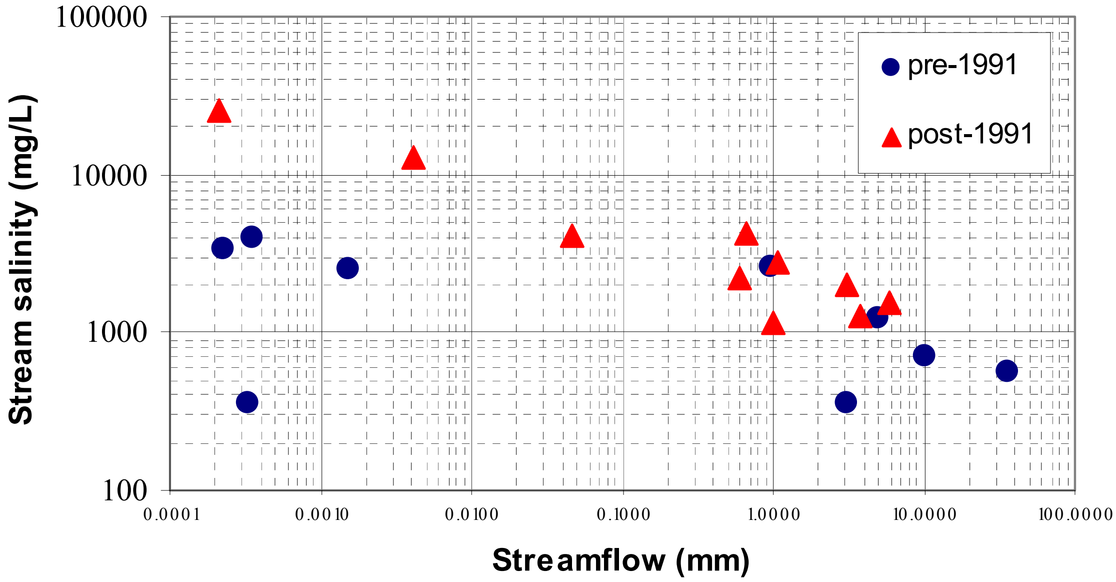

Figure 18. Flow weighted mean salinity vs. inflow to Toolibin Lake. Salinity of the inflow ranges from ~ 240 mg/L to ~ 26,000mg/L.

Surface water trend analyses (Figure 19) indicates an increase in salinity between 1984 to 1992, then a plateau at the salinity of 1,800 mg/L. A further upward trend after 1995 to ~ 2,000mg/L is also shown. This analysis suggests inflow salinity has increased by ~ 1,000 mg/L over the 20 years of inflow records, indicating significant saline groundwater seepage into the surface water.

The increase in surface water salinity during a period of little significant rise of watertable in the shallow bores in the valley flats and the yearly average unity value for the O/I of salt ratio from the catchment, needs to be explained. One hypothesis may be due to the impact of evaporation of shallow groundwater. These characteristics can only be explained if the rate of water removal from shallow groundwater were equal to or slightly less than the rate of watertable rise within the valley flats. This causes substantial increase in salt storage in the valley flats until the catchment reaches hydraulic equilibrium. The salt O/I ratio at that time is expected to substantially increase due to the leaching of the salt from the soil profiles. The increased seepage rate of saline groundwater means the increase of the salinity of the inflow into Toolibin Lake to values that reflect higher O/I salt ratio from the catchment.

Additionally, small increases in groundwater levels, and an expansion in the area of salinity may have resulted in the doubling of surface water salinity observed. However observations that O/I has remained near unity, suggests internal redistribution (within the soil profile or in detention stores) is driving the observed response. Evaporation driven process of salt mobilisation, detention and remobilisation at Toolibin, contrast with baseflow driven processes described by Bowman and Ruprecht (1999).

The results of surface water and groundwater trend analysis delineate the dominant hydrological processes occurring within the catchment. Water and salt balance calculations for the lake however can highlight the long-term impact of these processes on the salinity level of water within Toolibin Lake. Further, the water and salt balance can provide a tool to aid the evaluation of options for hydrological management of the lake.

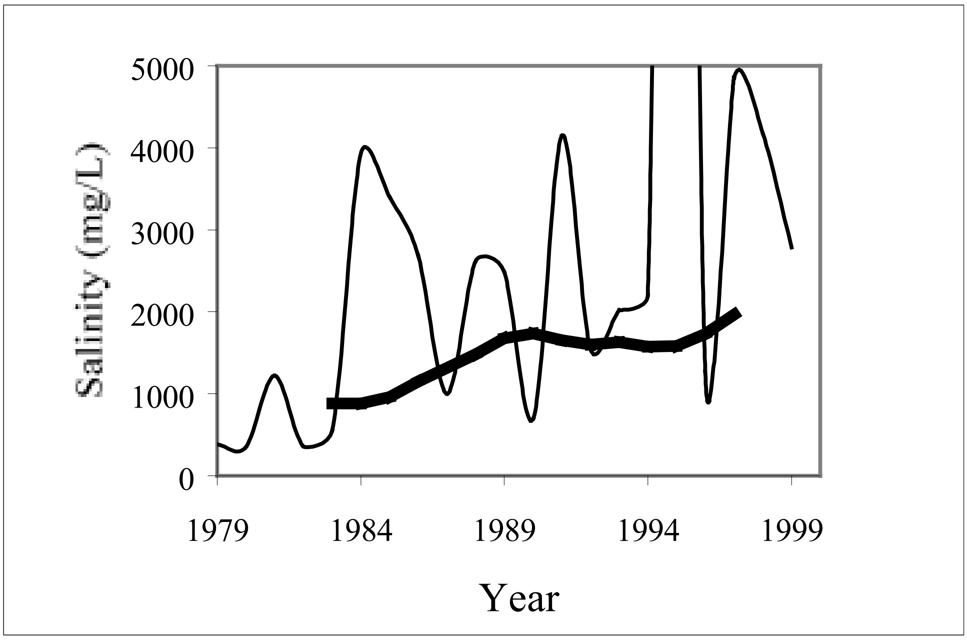

Figure 19. Surface water salinity trend. Note that salinity of the inflow to Toolibin Lake increased by ~ 1,000mg/L over 20 years.

6.3 Water and salt balance of Toolibin Lake

A model was used to evaluate the water and salt balance of Toolibin Lake. The adaptable spreadsheet for water and salt balance calculation (Barie & Peck, 2003) was developed to aid the evaluation of proposals to drain saline agricultural land into a lake. The model is based on a volumetric water balance equation, which used monthly time-step for data over the period January 1980 to December 2000 (the record used for surface water trend analysis) and a solute mass balance equation. The volume of water and mass of solute in the lake at the beginning of the simulation period should be known, but initial conditions may be guessed with some loss of accuracy during an initial period.

The model was calibrated to water levels measured in Toolibin Lake by manually varying values of outflow, channel hydraulics, lake-bed conductance and pan evaporation factor. Figure 20 shows measured and modelled water levels for a stream flow scaling factor of 1.15, a pan evaporation factor of 0.80, and lake bed conductance of $3x10^{-10}d^{-1}$.

As noted above, during several periods measured water levels were constant for significant intervals of time. As the elevation of the bed of the lake has not changed, the difference between modelled and measured water levels during these periods is a reflection of error in the reported elevation of the lake bottom (and therefore of the volume – elevation – water level relationships) and/or measurement error.

The model closely reproduces water levels in the lake during the period mid-1981 to early 1983 following periods of significant inflow. There is a poorer match of modelled and measured water levels following the 1983 flood event. The shape of the peak during the period of outflow (August 1983 to February 1984) is quite sensitive to assumed characteristics of the outflow channel. There is a very good match of modelled to measured water levels during the period of falling water levels after outflow ceased.

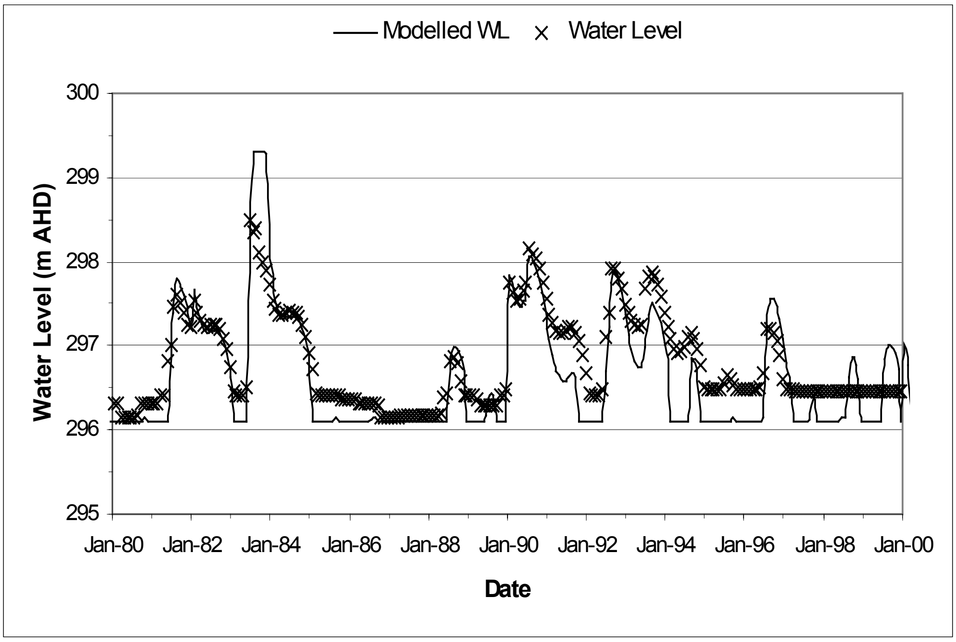

Figure 20. Measured and Modelled Water Levels in Toolibin Lake.

As presently formulated, the water balance equation in this model neglects runoff from areas of the bed of the lake that are not inundated with water, but it can be argued that any such input can be included in the stream flow scaling factor. There may be small losses of water by infiltration into the bed of the lake, as the lake expands to inundate peripheral areas, but it can be argued again that these can be included in the lake bed seepage term.

The results of the model suggest that the stream inflow and evaporation dominate the water balance of Toolibin Lake. Over the 21-year period January 1980 to December 2000, inflow and an additional 15% allowance for ungauged catchment totalled $3.58x10^7$ m^3. The estimated rainfall on the lake was $6.78x10^6$ m^3. The total evaporation and outflow from the lake, were $2.79x10^7$ m^3 and $1.5x10^7$ m^3 respectively and seepage through the lake floor totalled around 90 m^3 (negligible relative to other terms in the water balance of Toolibin Lake).

Stream outflow from Toolibin Lake is intermittent. The water balance model indicates outflow in only 10% of months during the 21-year period of simulation. The solute load in stream flow to Toolibin Lake is the dominant input. This model provides no indication that groundwater from aquifers beneath the lake makes any measurable contribution to either the water balance or the salt balance of Toolibin Lake. This is largely consistent with the observation that the majority of the lake floor has remained 'fresh' and that degradation is more recent (1980s onwards) and confined to the western area. It is also consistent with the observation that discharge rates are very low and salt is increasing in the root zone and is not being flushed.

The evaporation loss accounts for 77% of water budget of the lake leading to a four fold increase in salinity from an average of ~ 1000 mg/L (inflow salinity) to ~ 4000 mg/L. The salt balance calculated from the salinity of inflow and rainfall (7.2 mg/L, Hingston and Gailitis, 1976) incorporating seepage between the lake and an underlying aquifer suggests a substantial increase in the salt load in the soil profile of the lake.

The salt load in stream flow to Toolibin Lake is the dominant salt input. Over the 21-year period, total input in stream flow to the lake was 6.0×10^{10}g (60,000 T), including additional flow from ungauged areas, with rainfall on the lake surface contributing a further 4.88×10^{7}g (488 T). Loss of salt in stream flow from the lake totalled 2.5×10^{10}g (25,000 T) with the loss by seepage through the floor of the lake being 8×10^{5}g (0.8 T).

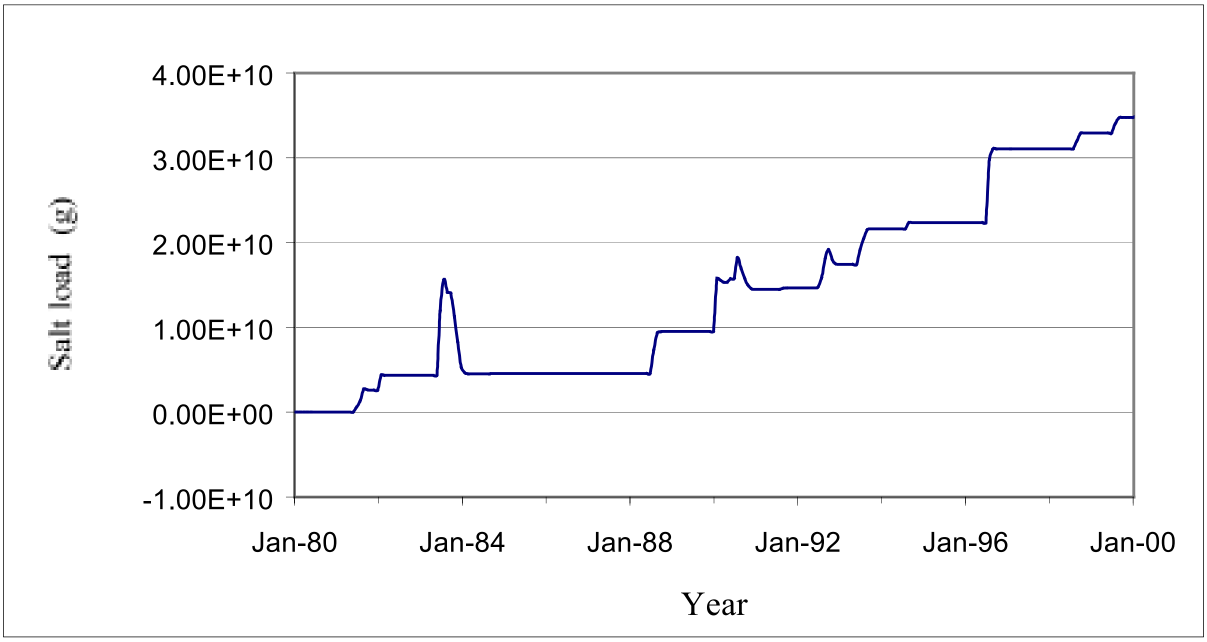

Figure 21. Modelled salt accumulation due to evaporation of surface water within Toolibin Lake.

The observed data suggests that the salt is flushed out of the lake, via the lake outlet or via seepage through the lake bed. However, when the lake level is lower than the outlet elevation accounting for approximately 2×10^{6} m^{3} of water, the flow stops and the water in the lake removed only by direct evaporation. The modelled salt accumulation due to this process suggests significant increase in salt storage within the lake bed (Figure 21). The calculated value of salt accumulation (~ 60,000 T to 25,000 T) amounts to less than 300,000 mg/L distributed within the upper eight meters (the thickness of the lacustrine sediments) of lake bed.

7. Catchment water balance

The variation in hydraulic parameters of the aquifers of the Toolibin catchment controls the groundwater storage, velocity and the watertable configuration. They also control the recharge discharge regime and their interrelationship with the other components of the hydrological cycle.

Quantification of water inputs and outputs to a catchment requires estimation of catchment's water balance that describes hydrological regime. Assuming a catchment in which the surface water divides and groundwater divides coincide, and for which there are no external inflows or outflows of groundwater, the annual water balance equation would take the form:

$$P = Q + E + \Delta S_s + \Delta S_G$$

Where P is the precipitation, Q the runoff out of the catchment, E the evapotranspiration, ΔS_s the change in storage of surface water reservoir, and ΔS_G the change in storage of the groundwater reservoir during the annual period.

The application of the water balance equation is only an approximation of the hydrologic regime in the catchment. However, the equation clarifies many of the interactions between groundwater flow and other components of the hydrologic regime. Furthermore, the water balance calculation reveals the important processes that control groundwater discharge and the spread of shallow watertable in a catchment.

The current discharge area of the catchment (where the watertable is less than two metres below the surface and salinity is evident) extends over 8% of the Toolibin Lake catchment, mainly within the valley flats (Figure 22). However, the watertable, particularly in upper valleys and at the catchment boundary, is still rising. This clearly indicates that the catchment has not yet reached hydraulic equilibrium. If trends in water tables continue, more areas, particularly in the valley flats up gradient of the main drainage lines and along the smaller tributaries, may be affected by salinity in the future.

The area underlain by shallow water table (more than two metres) at hydraulic equilibrium was mapped using the hydrogeology, local flow systems, topography and water level trend analysis (George 1998). Based on these factors, a shallow watertable may eventually (more than 100 years) extend over ~ 24% of the catchment (George 1998, George et al, 2001). This prediction however only indicates the extent of the area underlain by shallow water table. The long-term impact of water level rise in the valley flats on surface water inflow into Toolibin Lake can only be determined by further modelling. The long term predicted groundwater simulations might also determine the impact of the various salinity management scenarios on the salinity of inflow into the Toolibin Lake.

7.1 Catchment salinity management scenarios

Numerical modelling was undertaken to determine the magnitude of components of the catchment water balance and predict the future distribution of the shallow watertable. The distribution of a shallow watertable was used as an analogue of areas predicted to become saline (discharge areas). The MAGIC model (Mauger, 1996) was used. A detailed description of the principles used in the hydrologic assessment of vegetation impact and numerical engine for the model is outlined in Mauger (1996).

Drilling results and airborne geophysical data delineated aquifers, geological structures and groundwater flow in the catchment (George, 1998). This information provided the data to map the current distribution of groundwater levels. Numerical modelling was then used to predict the temporal changes in watertable and flow dynamics due to the change in the type of vegetation cover and engineering work across the catchment.

The hydrogeology of the Toolibin catchment is treated as three primary layers in the MAGIC model (Mauger, 1996, George 1998, George and Dogramaci, 2000). At Toolibin these are the superficial deposits, saprolite, and the combination of saprock and palaeochannel system. The palaeochannel aquifer was also mapped, and modelled as an additional layer. The superficial sediments are largely unconfined while the saprock and the palaeochannel can vary from confined to semi-confined. The model reflects the multi-layered nature of the aquifers, with each of these hydrogeological units characterised by distinct hydraulic properties.

The steady state model covers 482 square kilometres, incorporating all the boreholes in the catchment. Each cell within the model is 100 metres square resulting in 270 rows and 425 columns. The catchment model consists of 114,750 cells. The topographical data was taken from digital elevation maps and then kriged to obtain a value for each node within the model area. The thickness of each hydro-geological layer was interpreted from the geological logs of the existing bores in the catchment. Recharge was calculated for each individual cell based on vegetation cover and interpolated moisture data (field saturated capacity). The assumption for the leaf area index and hydrological parameters is outlined in the command file used in running MAGIC model (Appendix 2).

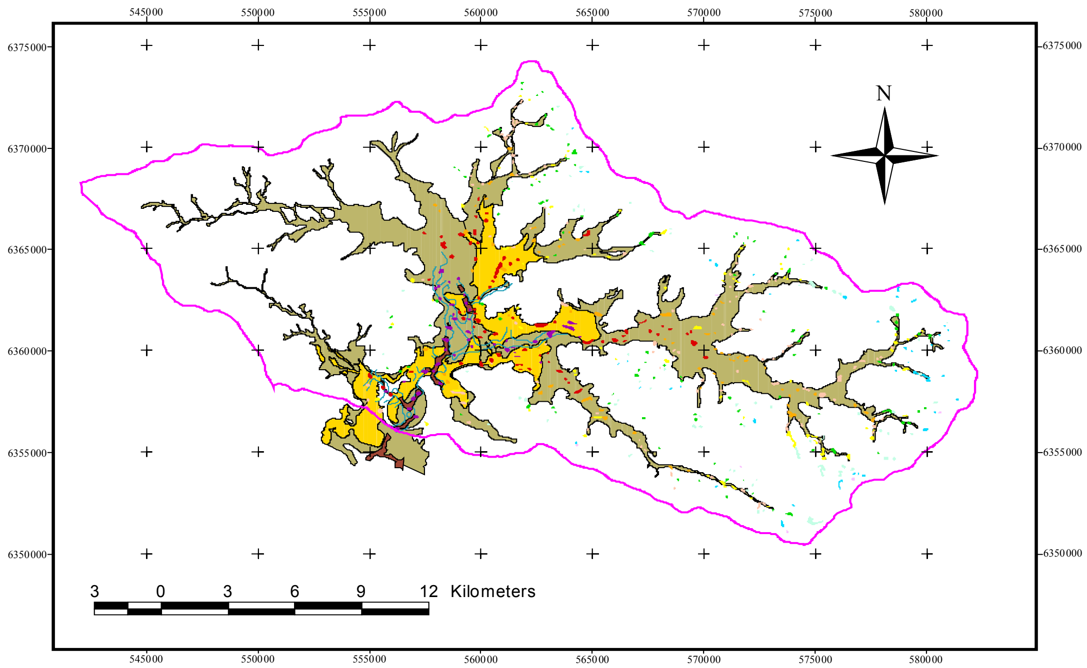

Figure 22. The current shallow watertable (more than two metres) and salinity area extending over 8% of the catchment (shown in yellow) superimposed on the predicted shallow watertable (equivalent saline area) at hydraulic equilibrium extending, over 24% of the catchment (shown in light green).

The inputs to the model are recharge due to precipitation and surface water inflow. The annual recharge rate was calculated using the estimated monthly field saturation index of the top layer. The top layers drain to lower layers and when recharge fills the bottom layers, and maximum lateral flow occurs (governed by gradient), excess water is routed downstream (surface flow).

An assessment of land use, groundwater salinity, topographic position and soil type was carried out to gain a first order approximation of recharge in the catchment. An initial assessment suggests that cleared areas, with relatively low groundwater salinity, are likely to have higher recharge rates. Groundwater in the catchment rim fits this category. However relatively high recharge rates may also occur on the valley flats (McFarlane et al., 1989).

The recharge rate for each cell was simulated using MAGIC (Mauger, 1996) over one 'verification' year of average rainfall. The year started at the beginning of August when soil water content is expected to be at its maximum. Assuming steady state condition, the soil water stored at the end of the 'average' year equals the initial 'stored water'. To estimate the soil water store for August, a run of the model commenced with saturated soil. Two years are then run in sequence with average rainfall to produce soil stores that can be used as the initial condition for the third year that represents the true 'average' year. The output of the monthly time step simulation is the potential recharge from each cell to the saprolite horizon. Potential recharge becomes actual recharge if there is sufficient capacity for deep groundwater flow away from the site, the capacity being determined by the transmissivity and hydraulic gradient at the site.

Two different values were used in the numerical modelling representing the upper and lower limits for average annual recharge rate. Values of 10 to 30 millimetres per year, equate to between 2% and 7% of the average rainfall in the catchment. These values are consistent with recharge values obtained for catchments in the Wheatbelt of Western Australia (Johnston, 1987; Nulsen, 1998).

The results of water balance calculation (Table 1) suggest that for an average rainfall of 392 millimetres per year and the average recharge rate of 10 millimetres per year (2.5%), the stream flow generated is four millimetres per year. The measured long-term average inflow into the Toolibin Lake is about 4.5 millimetres per year. A relatively higher recharge rate of 30 millimetres per year (7.6%) generates stream flow (runoff and groundwater seepage) of ~ 12 millimetres per year, three times higher than that in the previous scenario. The variation of the volume of surface inflow into Toolibin Lake shows that the higher flow rate of 12 millimetres per year is only observed for inflows of above average rainfall years (Figure 12).

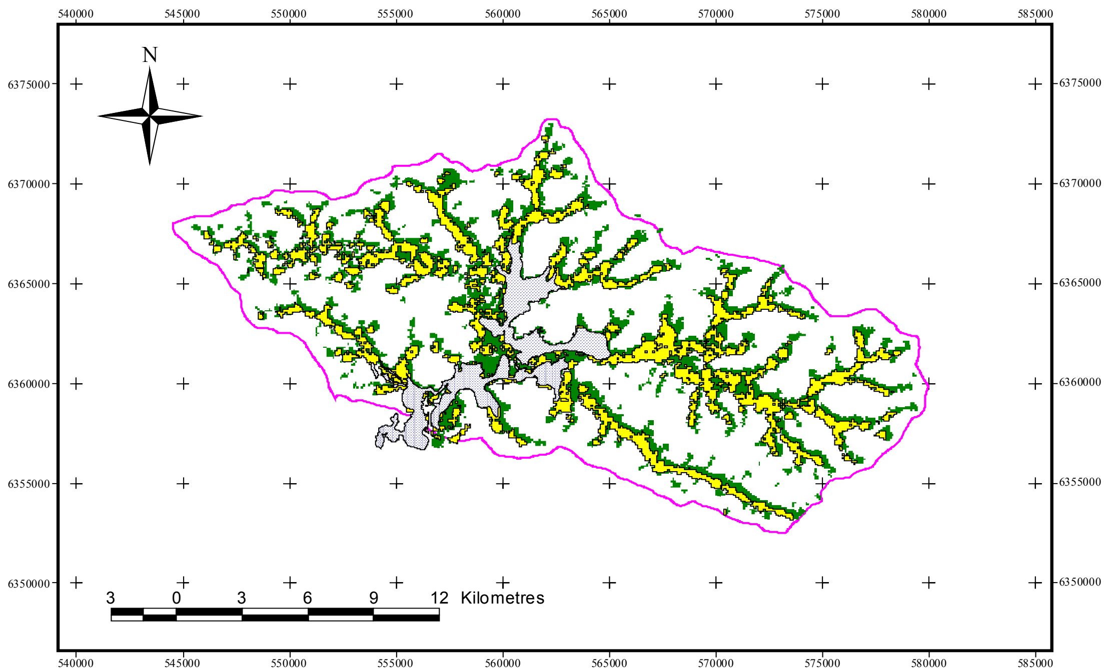

Figure 23. The extent of the shallow watertable for the recharge rates of 10 millimetres per year (yellow area) and 30 millimetres per year (green area). The area currently underlain by shallow watertable is shown in white.

The water balance calculations provide an estimate of the steady state inputs and outputs of the hydrological cycle. The calculations also highlight the relative importance of the dominant processes within the catchment. The low percent of stream flow generated in the catchment compared to rainfall, suggests that the majority of the rainfall is evaporated or transpired by vegetation. The low gradient of the landscape, particularly the micro depressions of the valley flats, provide an ideal environment for the runoff to accumulate and subsequently evaporate or recharge without contributing significantly to the total discharge from the catchment.

The area underlain by a shallow watertable for the average recharge rate of 10 millimetres per year (base case scenario) extends over 18% of the Toolibin catchment. It increases to ~ 32% for recharge rates of 30 millimetres per year (Figure 23; Table 2). This suggests that recharge rate has a direct impact on the extent and the configuration of shallow water at hydraulic equilibrium predicted by numerical modelling. The impact of variations of recharge rate on the extent of shallow watertable has been observed in other modelling studies at Toolibin (George et al., 2001) and in relatively high rainfall areas within the Collie catchment (Dogramaci et al., 2003).

Recharge scenarios	10mm per year	10mm per year	30mm per year	10mm per year
	Current	Hydraulic Equilibrium	Hydraulic Equilibrium	All forest
Stream Flow (mm)	4	7	12	0
Stream salinity (mg/L)	700	21000	110000	0
Shallow watertable (%)	8%	18%	32%	0%

Table 2. The extent of shallow water table for four different scenarios in the Toolibin catchment.

7.2 The effect of the shallow watertable expansion on surface water salinity

The current area underlain by shallow watertable covers only 8% of the catchment (Figure 23). The mean salinity of surface water comprising saline groundwater seepage and runoff generated from rainfall is ~ 1,800 mg/L. The salinity record over the last 20 years suggests that the surface water salinity has increased from 900 mg/L to 1,800 mg/L (Figure 19). However, the uncertainties in the measurement for low inflow volumes and the large variation in surface inflow preclude quantification of the magnitude of the trend in salinity. Nevertheless, anecdotal evidence and the deterioration of the lake environs support our analysis of a rising inflow salinity.

The predicted 10% expansion (i.e. more than doubling the area) of the shallow watertable at hydraulic equilibrium (recharge 10 millimetres) is likely to have an adverse impact on the salinity of surface water inflow to Toolibin Lake. Results from experimental catchments in high rainfall zones (~ 1,000 millimetres per year) indicate that the salinity of surface water has increased substantially when a significant portion of the landscape is underlain by shallow watertable (Johnston, 1987). This occurs because the rate of groundwater discharge is (at least seasonally) higher than the evaporation rate. Direct saline groundwater seepage (base flow) then mixes with surface runoff. Direct application of that model at Toolibin, with recharge rates of 10 millimetres per year and 30 millimetres per year, suggests that the salinity of the inflow could potentially rise to between 21,000 and 110,000 mg/L (Table 2).

However, because the salinity of surface inflow to the lake has only increased four fold since clearing, and the salt input/output ratio from the catchment is still close to one, it is expected that the increase in surface water salinity will take a relatively long time. Further, as the analysis of groundwater level trends from the bores in the valley flats suggests no significant trend over the last 10 years, the dominance of evaporation (allowing soil water salt storage) rather than base flow (direct injection to streams) might further mitigate the predicted increase. However, the continuous evaporation of groundwater will result in the accumulation of high levels of salt in the soil profile. The leaching of the accumulated salt will ultimately increase the salinity of the surface water and the inflow salinity into Toolibin Lake. Better management of the inflow to Toolibin Lake, and the re-diversion structure, might partly overcome the adverse impact of salinity increase. Furthermore, as a result of the variability of the rainfall and source areas of runoff, intense rainfall events will still generate high volumes of relatively low salinity water. This will create opportunities to capture and store fresher waters, even if the frequency of suitable events has diminished.

7.3 Recharge management options

One form of recharge management involves planting trees or other perennial vegetation on a large proportion of the catchment. The impact of tree planting on reducing the area underlain by a shallow watertable was modelled for four planting options. The analysis assumed a base case recharge scenario of 10 millimetres per year (Figure 23). Increased areas of land with a shallow watertable (32%) would result from use of 30 millimetres annual average recharge (Table 2).

Tree planting (at a density similar to a native woodlands) on the valley flats (~ 24% of the catchment) indicates that the area underlain by the shallow watertable would decreas from 18% to 14% (Figure 24; Table 3). This analysis assumes a non-negative interaction between the replanted vegetation and saline groundwater.

The impact of planting trees on the sand plain that covers 18% of the catchment, was similar to their impact in the valley flats. The reduction in the extent of area underlain by shallow watertable at the hydraulic equilibrium was ~ 4% of the catchment, with an identical watertable configuration to that of base case scenario (Figure 25).

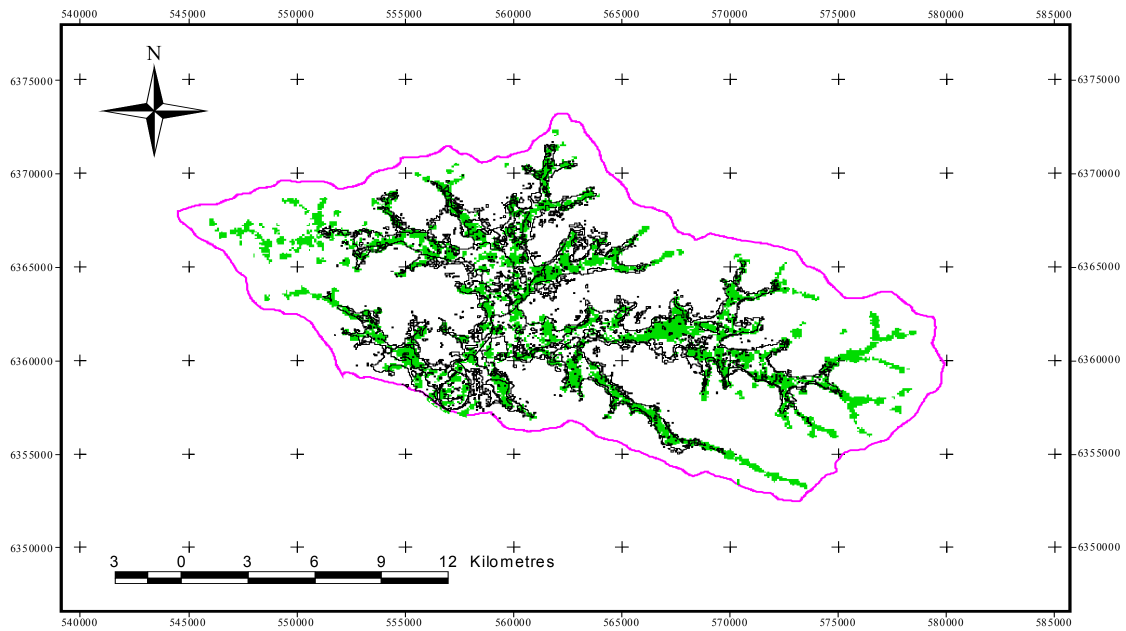

Figure 24. The tree planting area on the valley flats (24%) reduces the extent of shallow watertable at equilibrium from 18% to 14% of the catchment.

Testing various alley-planting designs across the catchment (Table 3) also shows similar results to that of sand plain plantings. Both result in a limited response of groundwater and equilibrium area of equivalent dry land salinity. For example, planting single rows of trees in 50 metre-spaced alleys has little or no impact if the modelled zone of 'no recharge' is maintained at or beneath the alley trees. However, when coupled with a widely distributed herbaceous perennial (like lucerne) in between tree belts, water tables are substantially reduced and there is effectively then no surface flow in the catchment. This model assumes perennials can be grown in a cropping and grazing system and that this system effectively eliminates recharge. In this case, lucerne is used as an analogue for other deep-rooted perennials. Suitable species are yet to be developed for soils in most upland areas of the catchment.

Similarly, planting 100% of the cleared areas within the catchment to a tree cover (at a density similar to native woodlands) shows that the area underlain by shallow watertable decreases from the current level of 8% to only 1% of the catchment (Figure 26). In this scenario, there is no surface water (runoff) due to the water uptake by the planted trees.

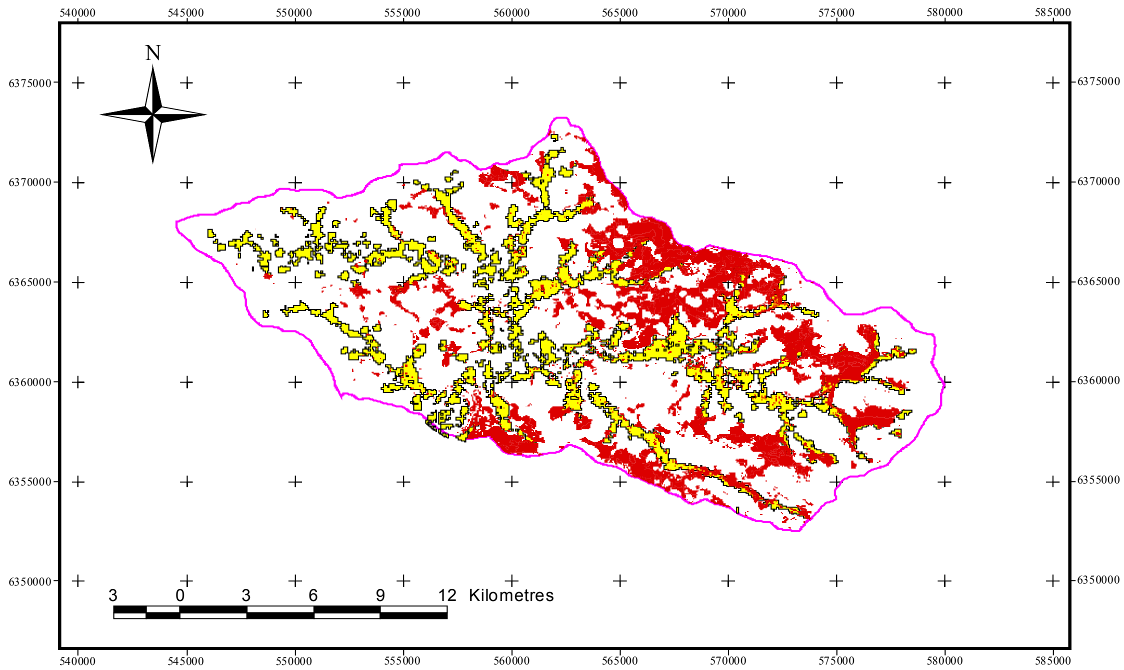

Figure 25. The planted sand plain area covering 18% of the catchment (dark colour) superimposed on the extent of shallow watertable at hydraulic equilibrium (light colour).

Tree planting scenarios	Sand-plain	100m Alleys covering the catchment	Valley Flats	50m Alleys covering the catchment	50m Alleys + lucerne
Stream Flow (mm)	3	1	3	NA?	0
Stream salinity (mg/l)	48397	293584	45570	NA?	0
Shallow Water table (%)	15%	18%	14%	18%	1%

Table 3. The extent of shallow water table for three different tree-planting scenarios in the Toolibin catchment. By comparison, the steady state, base case analysis suggests 18% of the catchment will eventually have a shallow watertable.

The results of the MAGIC modelling suggest that conventional recharge management options have a minimum impact on the reduction of the extent of shallow watertable at hydraulic equilibrium. Only plantings that influence the majority of the catchment, change the extent of shallow water tables. The principal reason for this poor outcome is the low transmissivity and watertable gradient, which restricts the lateral flow of recharge at rates greater than about one millimetre annual averaged recharge.

George et al (2001) obtained similar results on the likely impact of recharge reduction using Flowtube modelling. The Flowtube model is two-dimensional, cross sectional transient groundwater model (Dawes et al., 2000). In this model it is assumed that the groundwater body can be separated into a suite of flow tubes which are the thickness of the regolith, and each representing a single flow line in a flow net. The results of this model in the Toolibin catchment show that the long-term salinity risk responds only to significant reductions in recharge (large scale deep-rooted perennial planting).

Importantly however, George et al (2001) showed that even relatively low levels of intervention (local scale planting) could buy a significant amount of time before salinity impacts are fully realised. For example, modelling showed that alleys consisting of two row belts of oil mallees spaced 50 metres apart, could buy up to 40 to 60 years if established over the majority of the non-saline areas within the catchment. These treatments do not, however, significantly alter outcomes in the very long term (less than 100 years). Like MAGIC, Flowtube modelling also indicates that a large reduction in recharge is needed to produce a relatively small reduction in the extent of shallow watertable in the catchment. Finally, although these numerical models namely 'Flowtube' used by George et al., (2001) and MAGIC used in this study are different in terms of conceptualising catchment hydrogeology, and groundwater simulation, their results are remarkably similar.

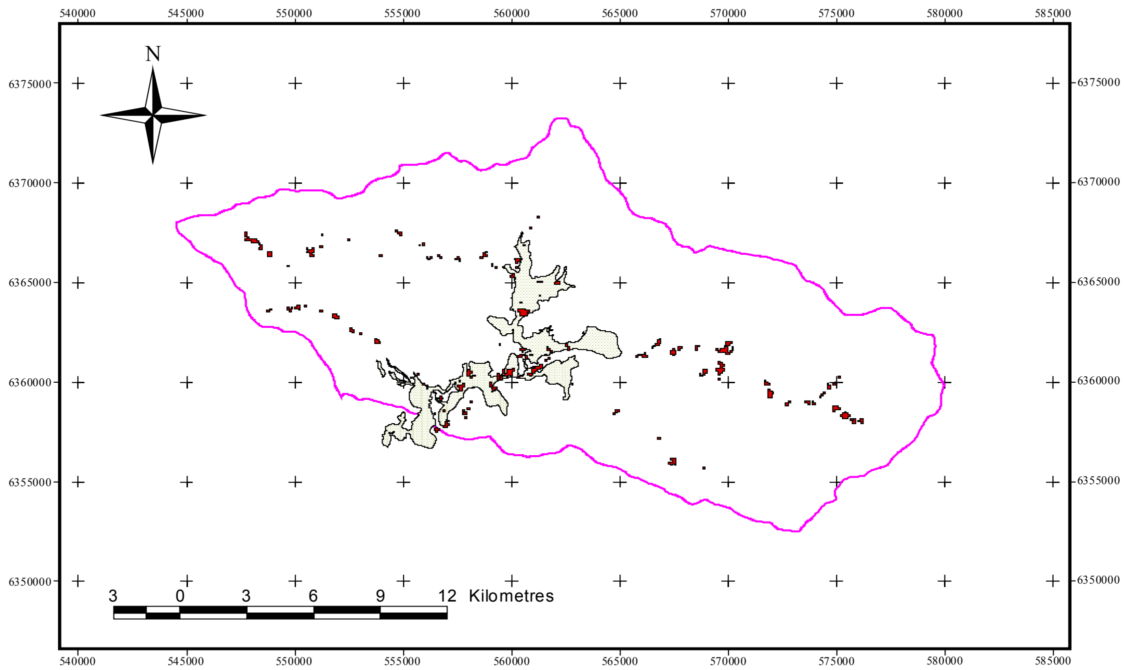

Figure 26. The predicted area underlain by a shallow watertable for 100% tree planting scenario (dark colour). The current area underlain by a shallow watertable (light colour) is also shown for comparison. Residual areas (red spots) are an artefact of the MAGIC modelling.

7.4 Impact of Groundwater Pumping

A MODFLOW groundwater model was developed using the hydraulic properties from the pumping tests to evaluate various pumping scenarios, and their impact on groundwater level beneath Toolibin Lake (SKM, 2000). The model was calibrated against seven years of monitoring records.

Eight scenarios, presented in terms of percentage of the lake area with a watertable more than two metres, were used to test the sensitivity of the model. As expected the results suggested that the model is sensitive to variations in hydraulic conductivity. The results of the modelling of various pumping scenarios suggest that without groundwater pumping, ground waters remain within two metres of the surface and the lake discharges saline groundwater. Continuous pumping however, about 1,000 kL/d for 4,000 days results in draw down of about two metres across 82% of the lake floor. At these rates, the watertable is lowered by one metre over the entire lake within three years.

The modelling of the impact of groundwater pumping on the salinity of Toolibin Lake was carried out by SKM (2000). Eight scenarios, presented in terms of percentage of the lake area with a watertable more than two metres, were used to test the sensitivity of the model. As expected the results suggest that the model is sensitive to variations in hydraulic conductivity. Increasing the hydraulic conductivity (Kx, Ky) reduces the impact to 40% of the lake, while reducing it increases the protected area to 90%. Changing the vertical hydraulic conductivity results in the protected area changing from 67 to 88%. Increasing and decreasing hydraulic conductivities to the east of the lake (lunettes) does not significantly change the predicted zone of impact (82 to 83%). Neither does reducing the connectivity of the palaeochannel and saprolite aquifers (83%) or reducing the hydraulic conductivity (Kx, Ky) of the dykes, which exist in the weathered zones (83%).

The results of the modelling suggests that the groundwater under the lake would remain within two metres of the surface and the lake discharges saline groundwater in the absence of groundwater pumping. In non-continuous pumping scenario (pumps are off when the lake is full); the results are similar to continuous pumping scenario, except that there is a lagged recovery. The last scenario looked at the frequency of filling, and found that a draw down of greater than two metres would be sustained over 82% of the lake by continuous pumping under the current filling regime.

Groundwater pumping systems have been sequentially implemented since 1996 when the first series of AirWell™ driven pumps were established on the western side of Toolibin Lake. This system operates with a combined well yield of approximately 200 kL per day. Submersible pumps in the palaeochannel to the east have been operating since April 1991 although periods of operating time has been lost while iron bacteria and related issues have been rectified. Since late 2002, the yield from two palaeochannel and single saprolite bore was 500 kL per day. Combined, the east and west systems deliver approximately 700 kL per day to Lake Taarblin evaporation area. The final system established at Toolibin approximates that modelled by SKM (2000). The major difference is that the expected well yields are lower than modelled (1,000 kL per day).

An assessment of the preliminary impact of the well field has been made difficult by the low rainfall experienced in the winters of 2001 and 2002. However, in summary, piezometric levels within the palaeochannel have been lowered by greater than one metre up to one kilometre from the pumping bores. Similar reductions can be observed in some of the western observation wells, although the impact zone is closer to these bores (300 metres). The impact on water tables is more pronounced to the west where the aquifer is largely unconfined. To the east, where the lacustrine systems confine the aquifer, drawdown has commenced although the rate is much slower and the impact of climate needs to be assessed further.

The Toolibin Recovery Plan set a target for a watertable level of two to three metres below the ground surface. While piezometric levels have been lowered to this degree in some areas, it is only possible to say that after 18 months, the modelled projections for the three to 10 year period appears consistent with early observations. Some qualitative evidence of the impact of lowered water tables can be seen near the western shoreline (P9 and P10), where pumping has been undertaken for the greatest period of time. Here regeneration of Casuarina stands has occurred over an area of approximately 100 to 200 metres from the bores.

8. Conclusions

- The hydrogeology of the Toolibin catchment is typified by complex regolith and related aquifers. Three local-scaled aquifers exist. At the weathering front, saprock ranges in thickness from 0.9 to eight metres and forms a well-connected groundwater flow system with a relatively high hydraulic conductivity. The overlying deeply weathered saprolite semi-confines the basement aquifer and has an average thickness of 13 metres. The upper layers of alluvium and colluvium consist of sand (more than two metre; sand plain hillsides) and sands and clays (more than five to 12 metres, valley alluvium) and its thickness is highly variable. Deeper sequences are only found in the lower catchment, e.g. within the western edge of Toolibin Lake.

- The most important 'intermediate-scaled' groundwater flow systems is a highly transmissive palaeochannel located at the eastern margin of Toolibin Lake extending ~ five kilometres up gradient northeast of the catchment, and an unknown distance downstream (Taarblin Lake). These alluvial channel sediments predominantly consist of fine sands from 20 to 30 metres in thickness. The sand particles are mainly sub-angular to sub-rounded quartz indicating short travel distance and rapid mode of deposition. Within Toolibin Lake the alluvial sediments are overlain by up to eight metres thick lacustrine fine clay that extends beneath the lunettes. It should be noted that the lateral extent of the palaeochannel is difficult to define.

- Vertical flows rather than lateral flows dominate groundwater movement. The lateral velocity of flow in groundwater in the catchment ranges between 0.001 and one metre per year. By contrast, vertical flux from the soil surface to the watertable may be of the order of tens of metres per year. Prior to clearing, groundwater may have taken tens of thousands of years to move to the outlet, while today, ground waters are rising to the surface and creating discharge areas in less than 100 years.

- The water levels in all areas of the catchment with a relatively deep watertable (less than five metres) are rising by approximately 0.1 to 0.5 metres per year. Ground waters in areas at risk in the upper catchment are between two and 20 metres below ground (e.g. Scriveners Soak catchment), suggesting equilibrium may take over 100 years to be fully achieved. By contrast, the average long-term rise in the shallow bores within the saline valley flats is close to zero, and now reflects seasonal rainfall patterns.

- The hypothesis used to explain the relatively stable water level patterns in the valley flats is that the rate of evaporation of groundwater is similar to the rate of water level rise. This process will, however, result in the accumulation of salt in the soil profile and thus potentially increase the severity of impact on vegetation. The increased load may lead to large episodic flows of salt if mobilised from saline areas.

- The results of Toolibin Lake water balance suggest that stream inflow and evaporation dominate the water balance, and the rate of seepage from the lake bed is negligible. When the lake level is lower than the outlet elevation the water in the lake is removed by evaporation. The modelled salt accumulation due to this process suggests significant increase in salt storage within the upper eight metres of lake bed. Additional salt derived from the shallow groundwater is also likely to be accumulating in the lake bed sediments.

- The current area underlain by shallow water tables extends over ~ 8% of the catchment. This area is expected to increase to between ~18% and 34% of the catchment at hydraulic equilibrium. The final area is dependent on the volume of accumulated recharge relative to the annual discharge rate. It must be noted that the modelled areas are not equivalent to the saline area. Actual soil salinity levels are the result of more complex surface water groundwater interactions.

- The short record for water level measurements and surface water inflow coupled with large variation and intensity of rainfall events precludes confident prediction of rates of salinity increase and particularly, accurate stream water salinity's at equilibrium. Our analysis concludes that salinity has increased by 1,000 mg/L over the 20 years of recording. The results of the numerical modelling of equilibrium levels suggest that if the saline groundwater seepage area is 18% of the catchment, annual averaged surface water salinity's may exceed 21,000 mg/L.

- The impact of recharge management options by tree planting on ~ 25% of the catchment is minimal in terms of the estimated area of the catchment that will be underlain by shallow water tables at equilibrium. However, recharge reductions (less than 30%) may reduce the rate of watertable rise and buy considerable amount of time in areas at risk. Conversely, if the majority of the catchment were planted the current area of shallow watertable might contract to 90% of current level. Only in this scenario will surface water salinity will be lower than the current level.

- Ensuring the survival of Toolibin Lake will depend on continued engineering-based management of surface water and the extraction of groundwater. In the longer term, plant based options may reduce the severity of impact and buy time until systems of recharge control can be adopted that are both practical and profitable.

9. Recommendations

The Toolibin Lake catchment is one of the most well studied and documented areas of the Western Australian Wheatbelt. The Toolibin catchment has become a testing ground or 'living laboratory' for the development, analysis and documentation of land and water management systems. This report summarised the hydrogeologic analyses and related assumptions conducted over the past 25 years, focussing on more recent findings (1998 to 2002). However there are several key research and development issues that are yet to be resolved.

The following recommendations are made regarding priority hydrology-based research and development within the catchment and at Toolibin Lake.

9.1 Toolibin catchment

Recharge manipulation based on current vegetation systems and adoption rates appears to be unable to prevent the development of shallow water tables. However modelling indicated that some planting configurations can reduce the rate of watertable rise and may buy a significant amount of time. Research and related demonstrations that test this should be undertake in the mid to upper reaches of valleys. Research should test the hypothesis that recharge could be limited, rates of watertable rise reduced and the degree to which both of these translate into the recovery or containment of saline land, assessed. The impact of oil mallee alleys and associated perennial or long season annuals should be specifically explored.

The second area of investigation requires the analysis of systems that could be called 'preventative engineering'. This would take the form of surface water control structures to lessen the risk of valley derived recharge and establishing groundwater pumping systems in areas with a moderate depth watertable (five metres). In the both cases, information on the extent of zones of protection and the period of protection afforded to such risk areas should be determined. Management of discharged waters would also form part of this evaluation.

9.2 Toolibin Lake

In 1995, the Recovery Plan for Toolibin Lake and related ecosystem was based on three priority actions:

a) Diversion of saline flows and flushing of Toolibin Lake;
b) groundwater pumping to reduce the impacts of shallow water tables and salt discharge; and
c) reduction of the long term trends in water tables and soil salinity within the catchment.

Our analysis of the current range of vegetation-based options and their effectiveness, suggests we should review options (a) and (b) again, because option (c) is unlikely to provide control in the short term.

Trends in surface inflow suggest a doubling of salinity levels in the past 20 years in response to the increased area of dry land salinity. It is also apparent that much of the stream flow is generated from areas that are not yet saline, and that an increase in the proportion of flow available to the lake, could be established by surface water engineering systems. In conjunction with such a surface water-based study, the current operational models for the separator and related structures should be reviewed.

The composite pumping systems have been operational for over 18 months, and a review of the effectiveness of the well field and its impact on water levels is required. This review should be undertaken in two parts. Firstly, an analysis of the monitoring and flow data is required to assess the impacts of the well-field and compare it to the earlier predictions by SKM (2000). This review should be undertaken in 12 months time (after winter 2003) and should include a detailed hydrograph analysis that takes account of the impact of climate variability.

Secondly, consideration should be given to a specific assessment of the impact on pumping on root-zone salinity, particularly given results of our salt balance study that concludes that there has been little salt derived from groundwater exported from the lake. Revisions to the groundwater model and projections should be made. Operational issues relating the bore yields, power demand and iron biofouling, should also be assessed. Analysis of the likely impact of both reduced and increased bore yields, or additional bores should be considered.

There is also a need to develop more sophisticated predictions of the future extent of salinity in two distinct areas. The first is related to the impact of climate. Groundwater monitoring indicates that the droughts of 2001 and 2002 have in some cases reduced the rate of watertable rise. Scenario modelling of a range of climate ‘forecasts’ should be undertaken using the scenarios modelled by MAGIC and Flowtube, to review the runoff and recharge estimates. A catchment water balance model should report these.

Surface and groundwater monitoring represents a very important component of these analyses being undertaken at Toolibin and in many ways is what separates this catchment from all others in the WA Wheatbelt. It is strongly recommended that the current-monitoring schedules be maintained and improved.

Finally, the relationship between the area underlain by a shallow watertable and the area of salinity should be established. This can be done in two ways, by comparing the area of a shallow watertable at present with the area saline (by use of soil salinity assessment), and by predictions in areas at risk. This work should also include an assessment of groundwater discharge rates and soil salinity changes forecast to be taking place in the unsaturated zone.

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11. Appendix 1

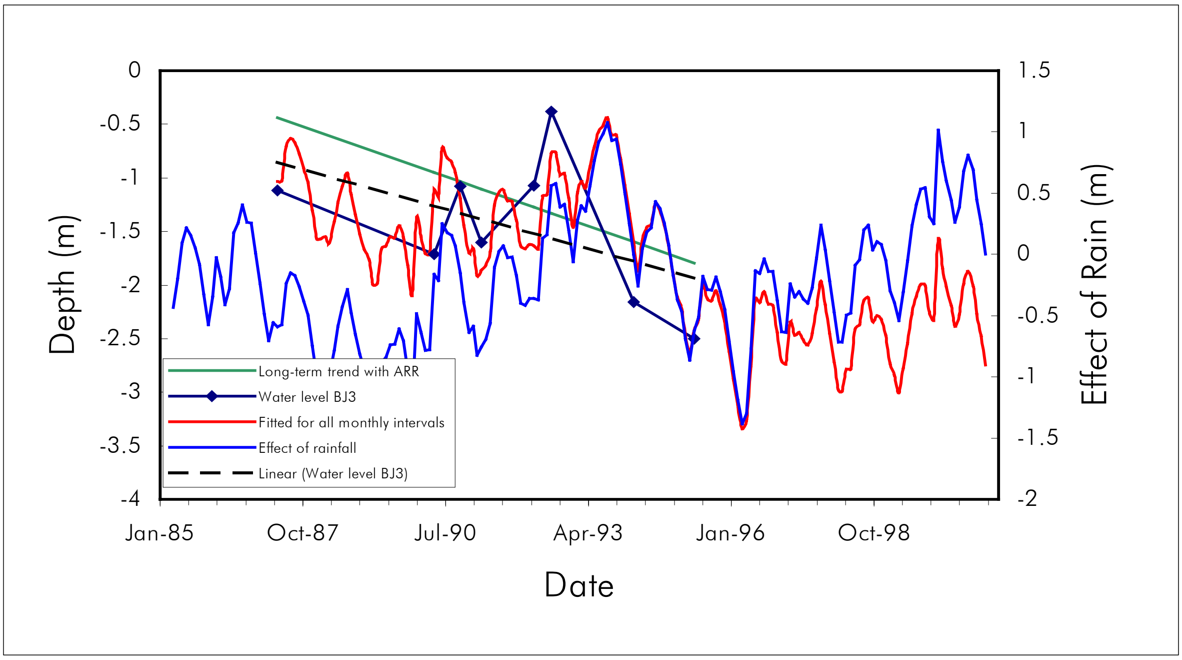

Figure 1. Water levels with accumulative annual residual rainfall for BJ3 (0 months delay).

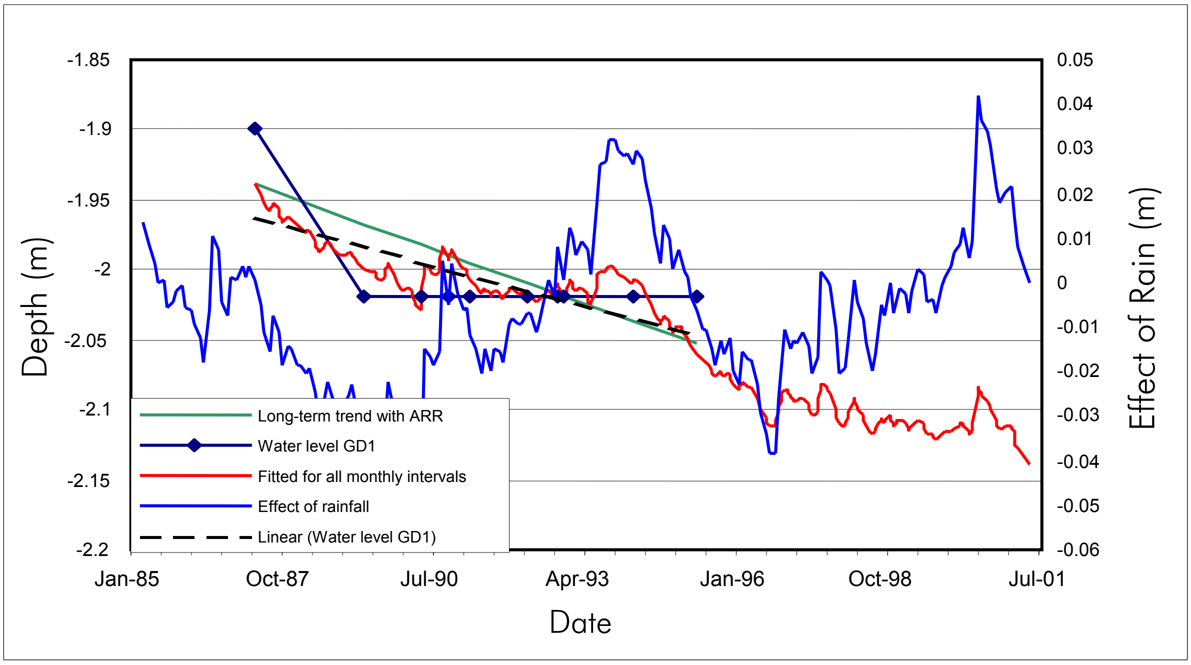

Figure 2. Water levels with accumulative monthly residual rainfall for GD1 with the data correction (five months delay).

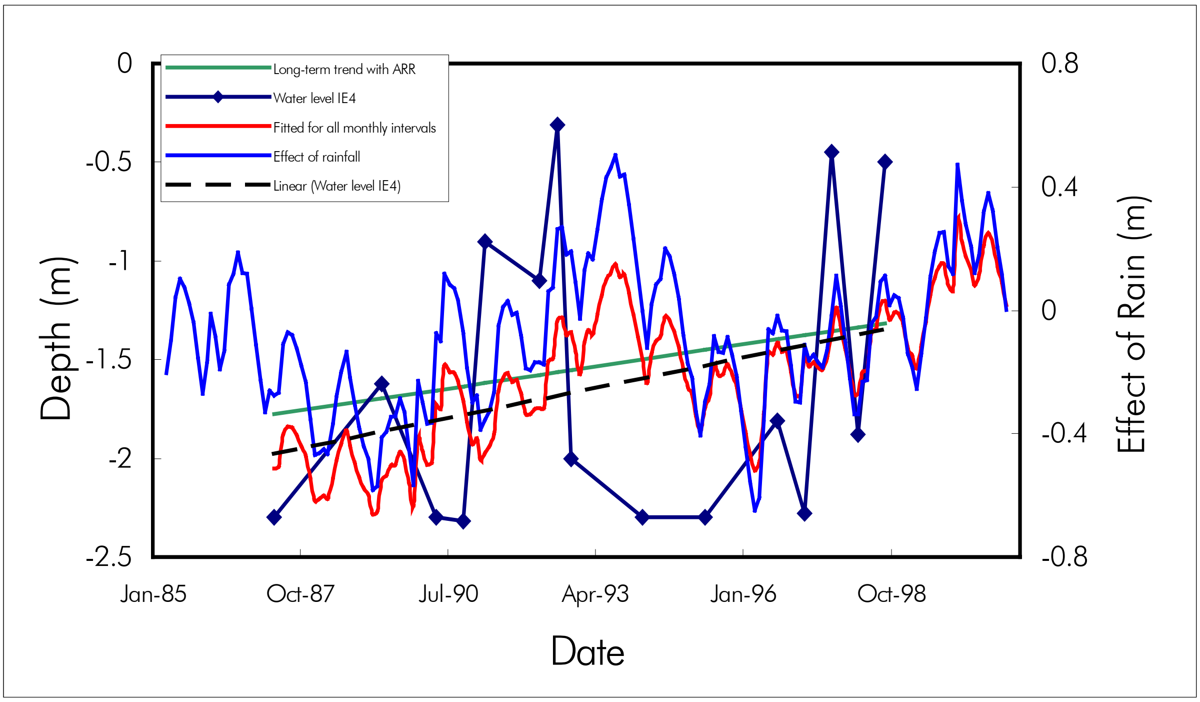

Figure 3. Water levels with accumulative annual residual rainfall for IE4 (0 months delay).

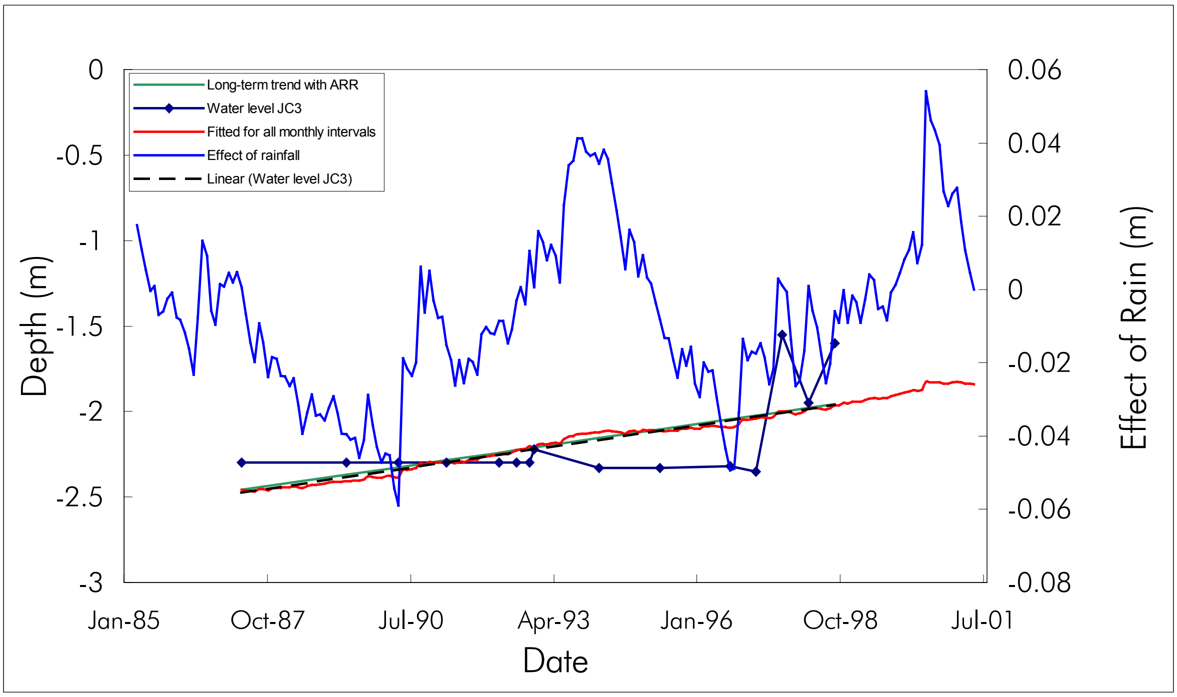

Figure 4. Water levels with accumulative monthly residual rainfall for JC3 (five months delay).

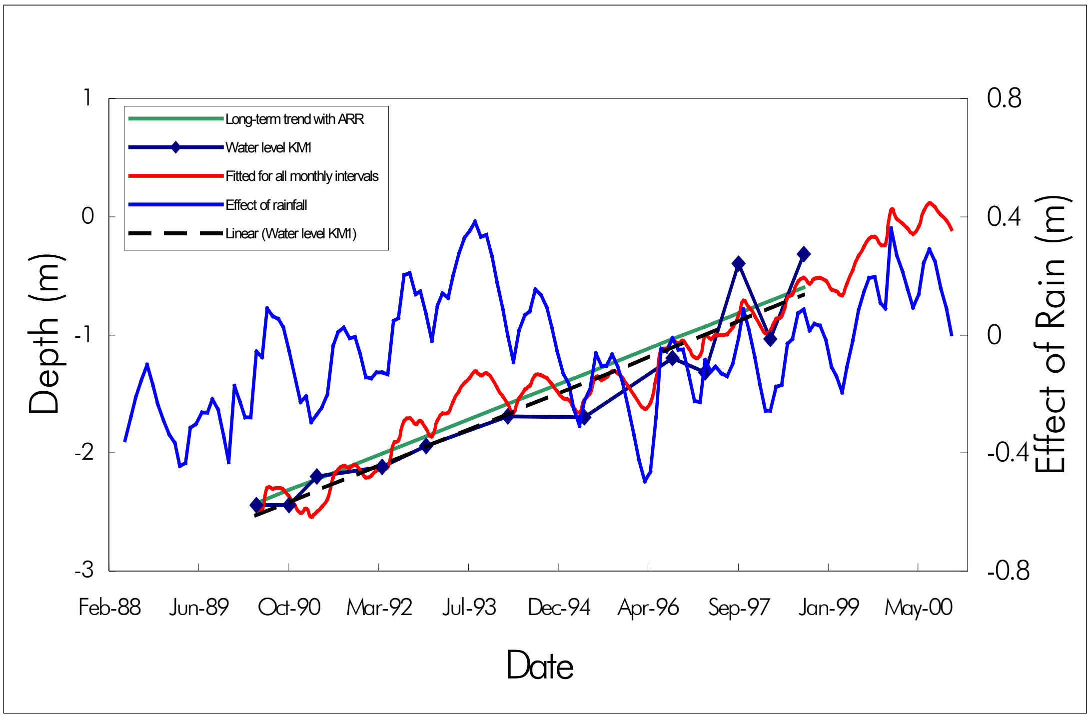

Figure 5. Water levels with accumulative annual residual rainfall for KM1 with the data correction (0 months delay).

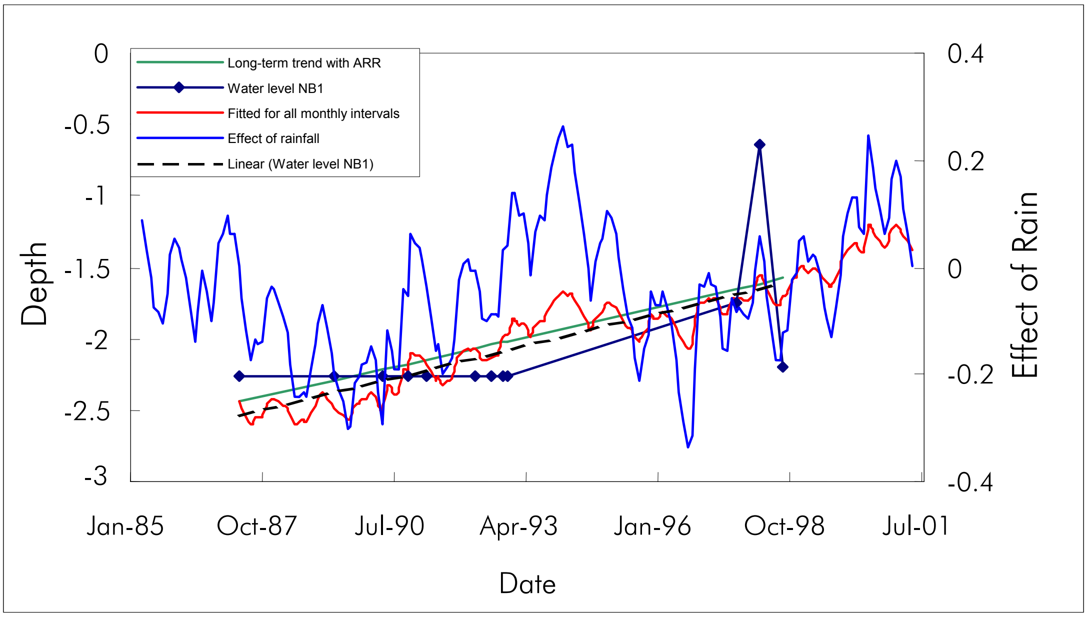

Figure 6. Water levels with accumulative annual residual rainfall for NB1 with the data correction (five months delay).

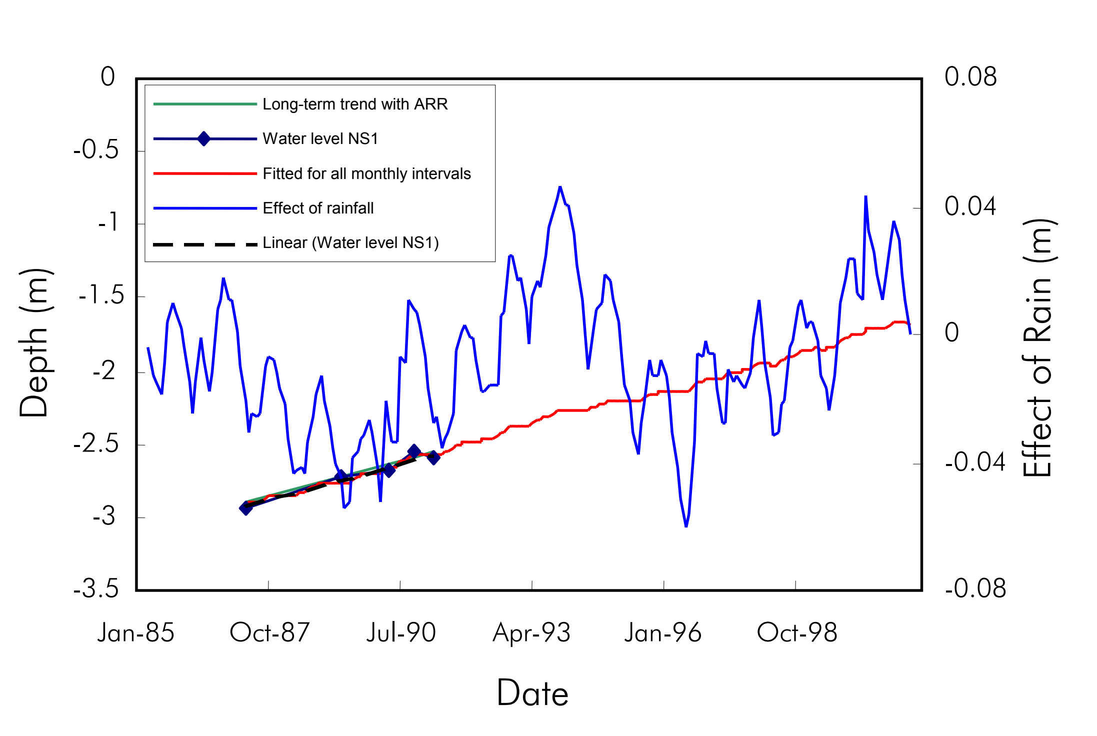

Figure 7. Water levels with accumulative annual residual rainfall for NS1 (three months delay).

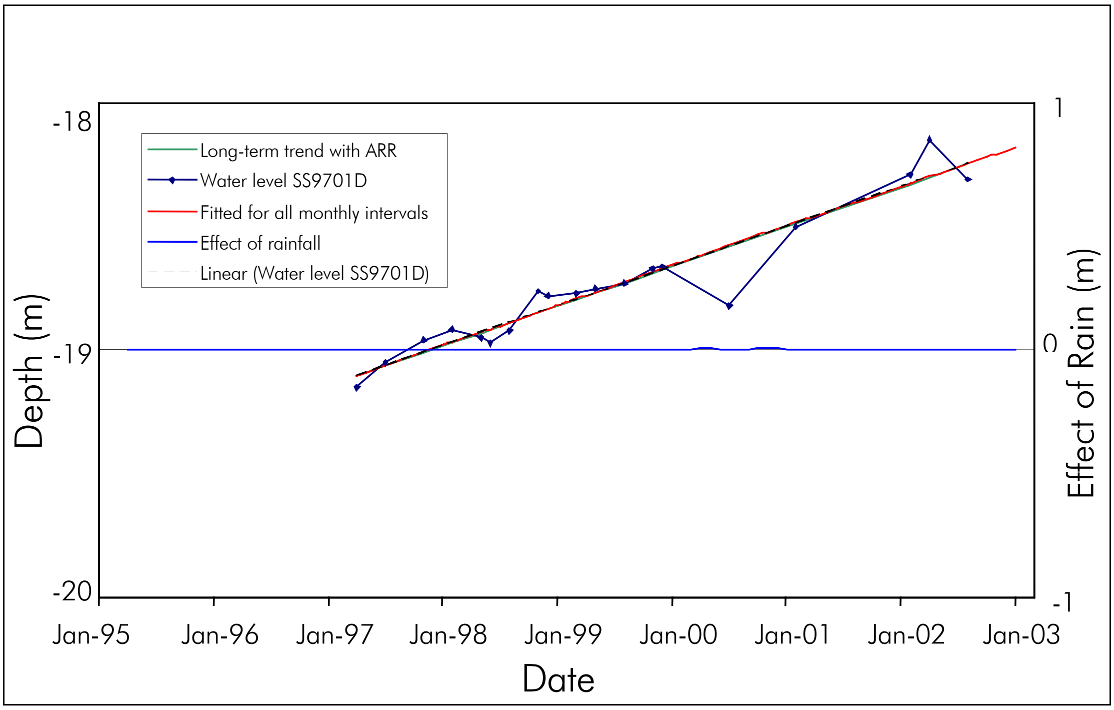

Figure 8. Water levels with accumulative monthly residual rainfall for SS9701D (four months delay).

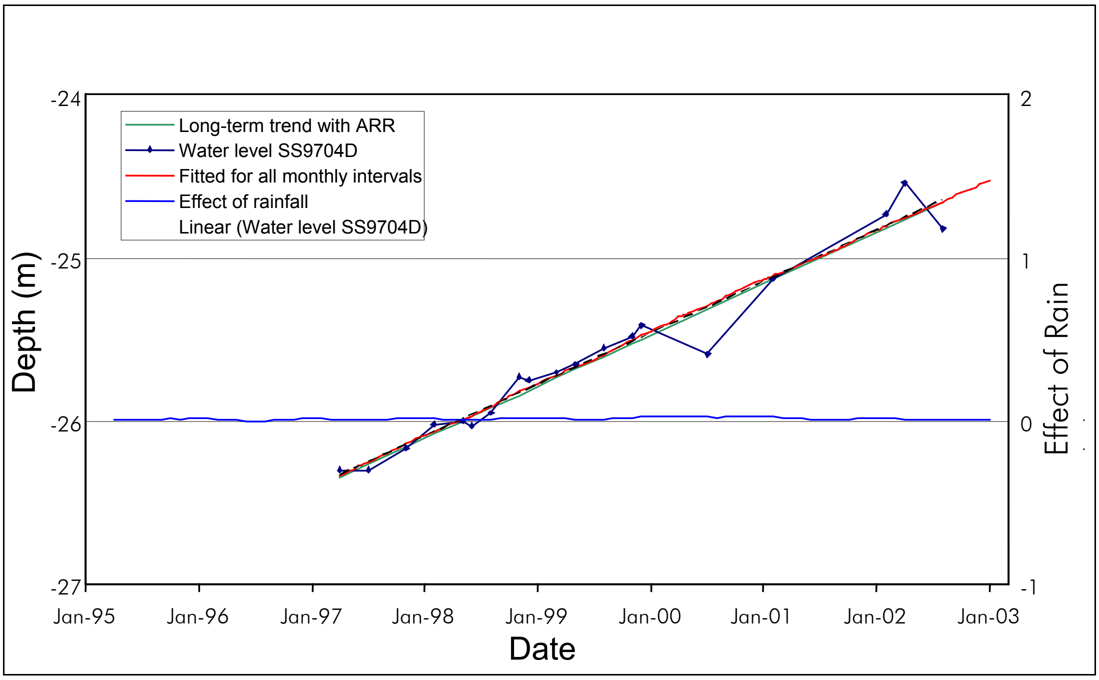

Figure 9. Water levels with accumulative monthly residual rainfall for SS9704D (four months delay).

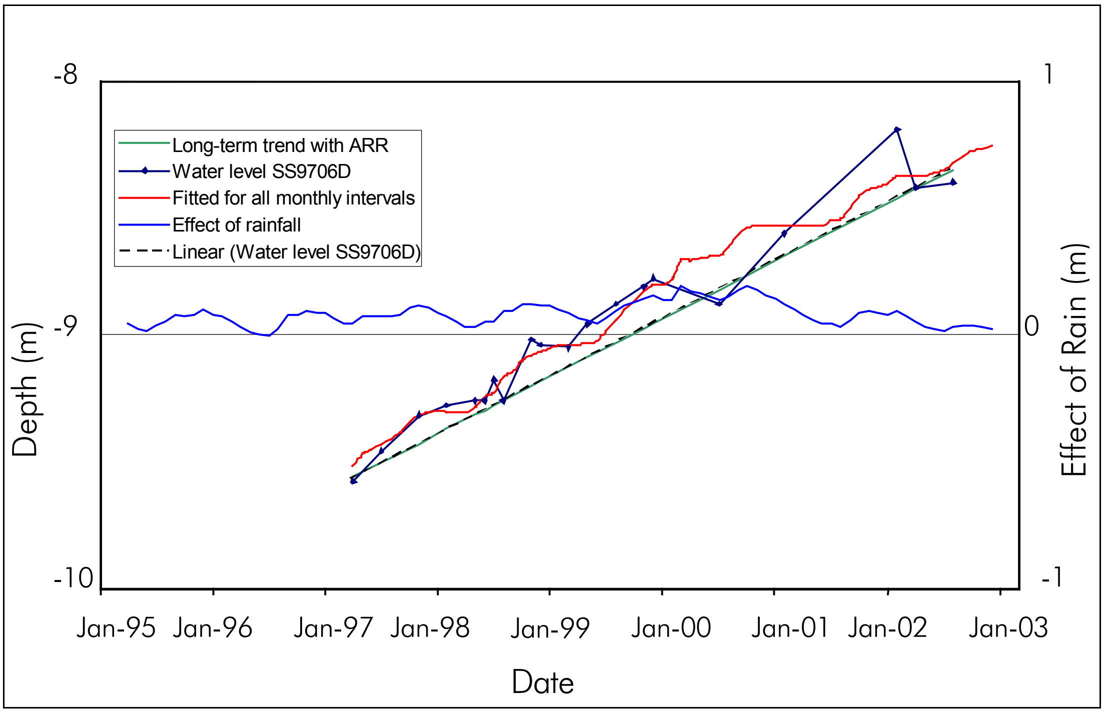

Figure 10. Water levels with accumulative monthly residual rainfall for SS9706D (three months delay).

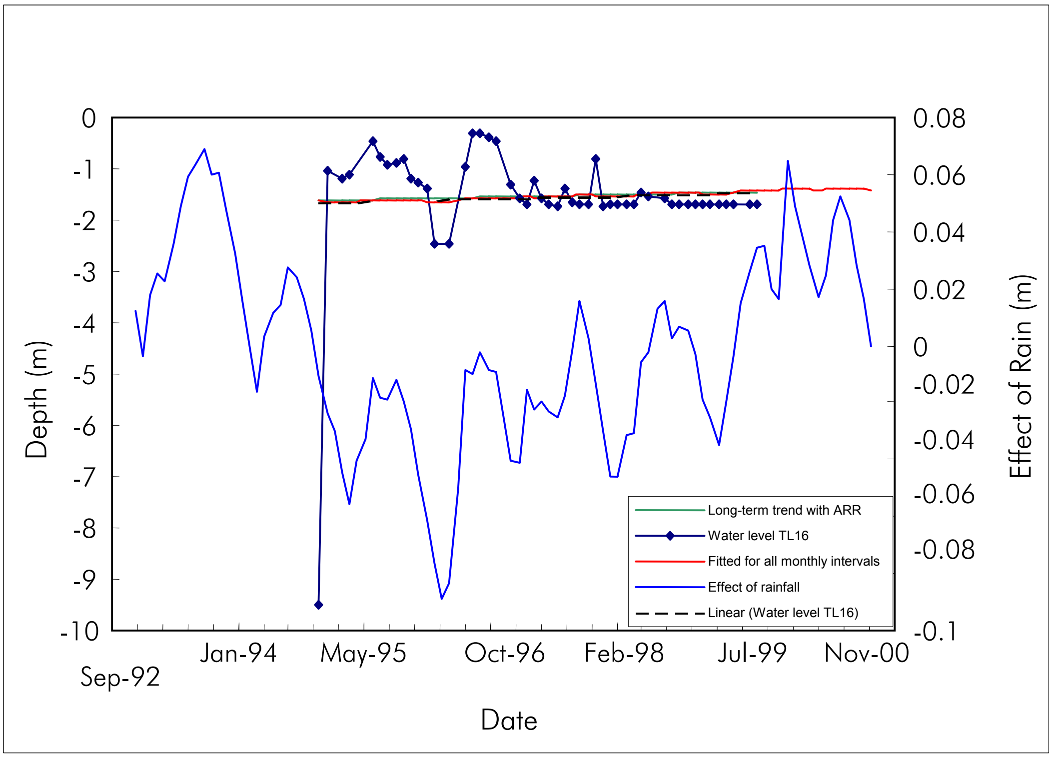

Figure 11. Water levels with accumulative annual residual rainfall for TL16 (0 months delay).

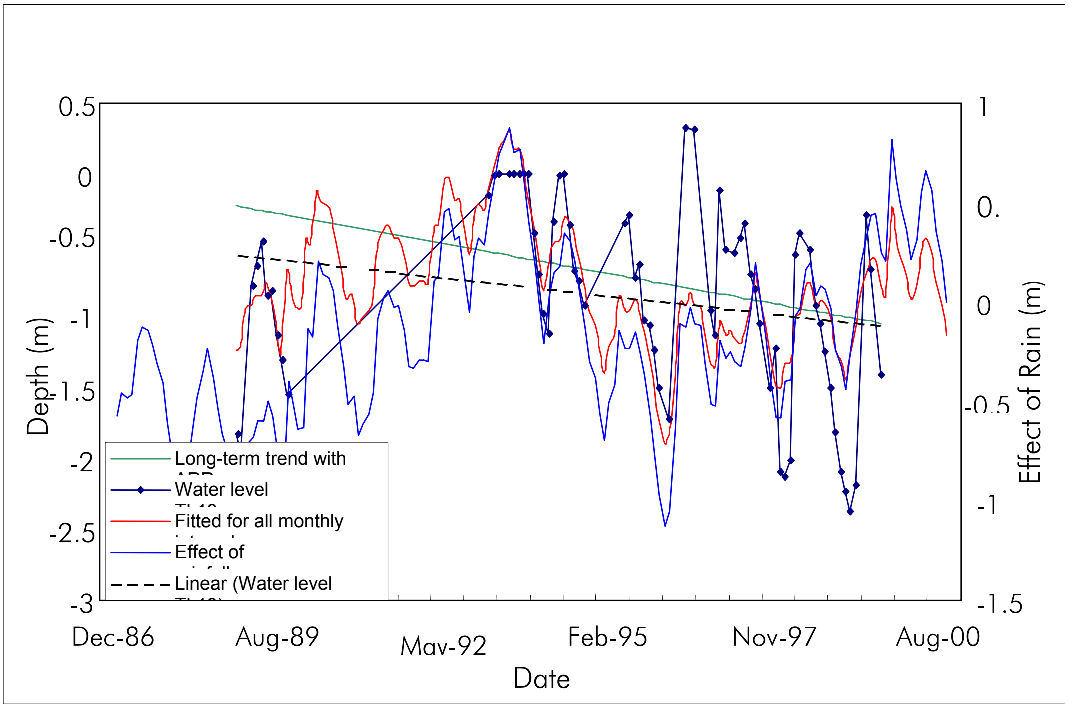

Figure 12. Water levels with accumulative annual residual rainfall for TL19 (0 months delay).

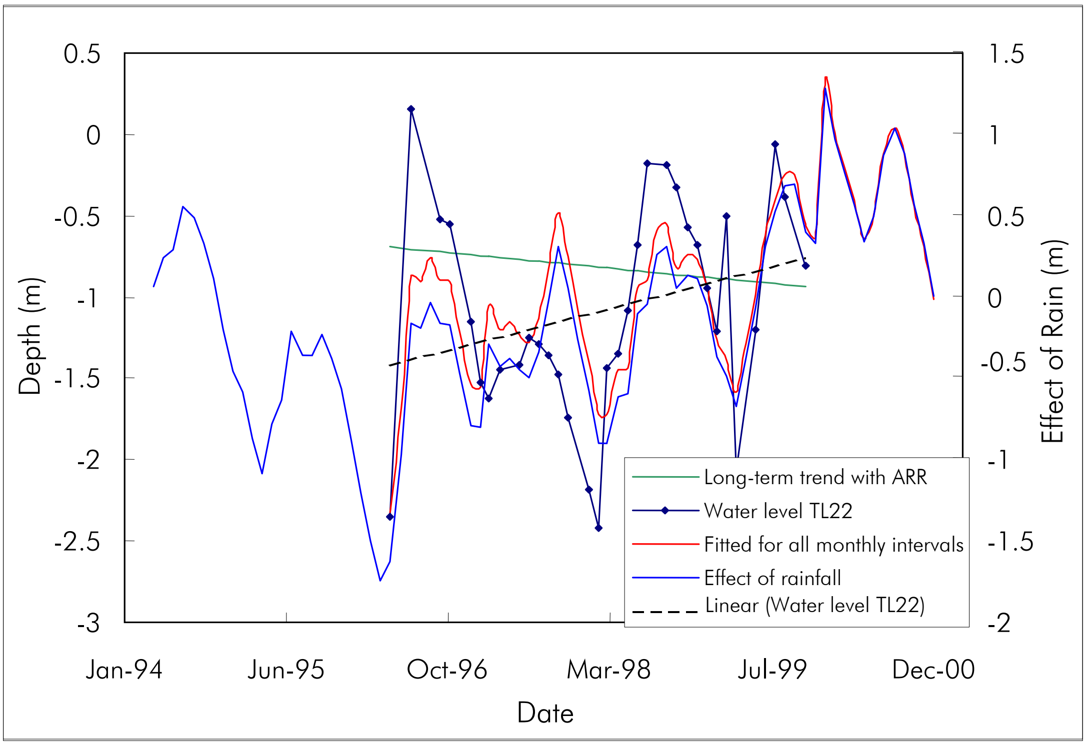

Figure 13. Water levels with accumulative annual residual rainfall for TL22 (0 months delay).

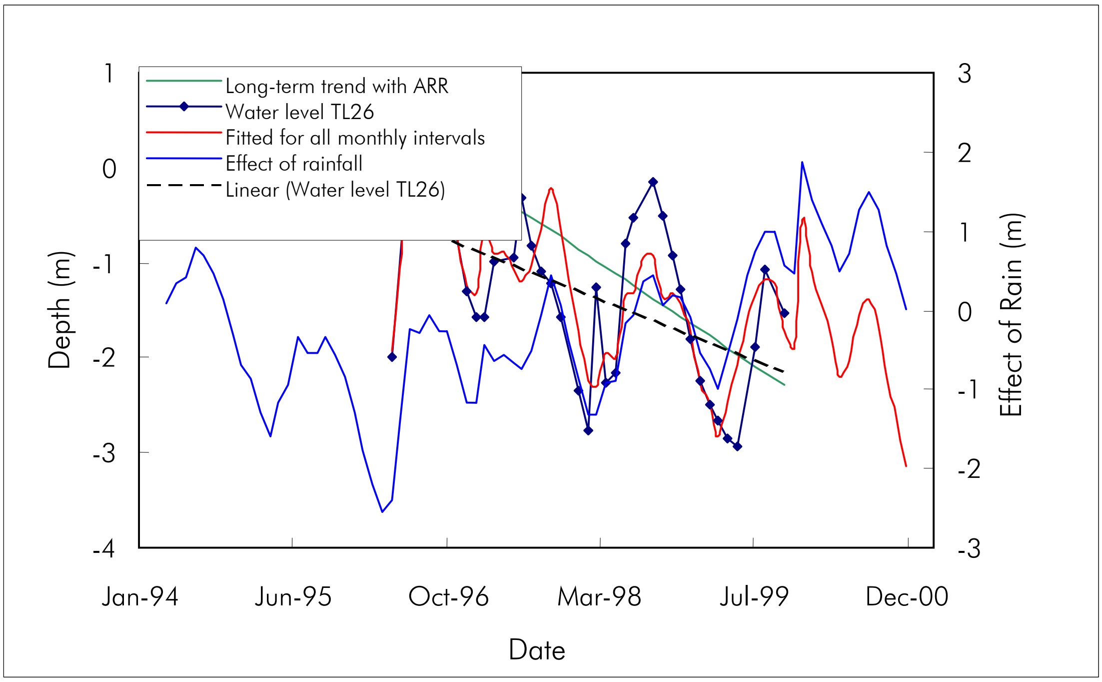

Figure 14. Water levels with accumulative annual residual rainfall for TL26 (0 months delay).

12. Appendix 2

Command file used to run the MAGIC model for Toolibin catchment

```
SET # 8
SET CELLSIZE 100
SET AREA CELLSIZE^2
SET FACTOR AREA/625
SET DRY 20*FACTOR
SET DPLAI 2.1  : LAI OF DEEP ROOTED PASTURE
SET CLAI 1.5   : LAI OF SHALLOW ROOTED PASTURE
SET DEPTH 1.5  : THICKNESS OF TOP LAYER (M)
SET K 30       : KSAT PERMEABILITY OF TOP LAYER (M/MTH)
SET KD 1442     : KSAT TRANSMISSIVITY OF BOTTOM LAYER (M/YEAR)=(.8*3+.1*15.5)*365*2
SET VC 2      : VERTICAL PERMEABILITY OF CLAY LAYER (m/yr)
SET POROSITY .2
SET WATERST (POROSITY*DEPTH*AREA)
SET DEEPVOL (AREA*.2*3)
SET CLAYVOL (AREA*.2*15.5)
SET MINCLAY (AREA*.2*10)
SET DGWSAL 20000 : MG/L SALINITY OF DEEP GROUNDWATER
SET CLAYINF M130 : MAX INFILTRATION RATE FROM TOP LAYER INTO UNSATURATED CLAY
SET CLAYTRANS 8 : MAX INF. RATE FROM TOP LAYER INTO SATURATED CLAY ENROUTE TO
BOTTOM LAYER
: vary g/w inflows about month average by rainfall scaled by GWSCALE
SET GWSCALE .7
: fraction of salt in transpired volumes that transfers to dry store=PRECIP
SET PRECIP 1
: fraction of lake evap to apply to runoff
SET EFRAC 1
SET BNDY M150
SET STORE1 M246
SET STORE2 M247
SET STOREC M268
SET ISTORE1 M461
SET ISTORE2 M462
SET TEMP1 M463
SET TEMP2 M464
SET MAXSTORE2 M279
SET ETTREE M248
SET LTREE M427
SET UTREE M426
SET ETAP M242
SET LAP M424
SET ETDRP M259
SET LDRP M425
SET UDRP M429
NAME(TEMP1,"TEMP_1 SCRATCH MAP","R4")
NAME(TEMP2,"TEMP_2 SCRATCH MAP","R4")
NAME(M421,"SCRATCH MAP","R4")
NAME(STORE1,"SHALLOW LAYER STORAGE #","R4")
NAME(STORE2,"DEEP LAYER STORAGE #","R4")
NAME(STOREC,"CLAY LAYER STORAGE #","R4")
NAME(ISTORE1,"START SHALLOW LAYER STORAGE #","R4")
```

```
NAME(ISTORE2,"START DEEP LAYER STORAGE #","R4")
NAME(MAXSTORE2,"MAX DEFICIT OF DEEP STORE #","R4")
NAME(ETTREE,"ACCUMULATED TREE ET #","R4")
NAME(ETAP,"ACCUMULATED ANNUAL PASTURE ET #","R4")
NAME(ETDRP,"ACCUMULATED DEEPROOTED PASTURE ET #","R4")
NAME(LTREE,"TREE ET LEFT #","R4")
NAME(LAP,"ANNUAL PASTURE ET LEFT #","R4")
NAME(LDRP,"DEEPROOTED PASTURE ET LEFT #","R4")
NAME(UTREE,"TREE ET USED #","R4")
NAME(UDRP,"DEEPROOTED PASTURE ET USED #","R4")
NAME( M447 ,"FINAL +VE DEEP DRAINAGE #")
NAME( m448 ,"negative throughflow #")
NAME( M269 ,"DEEP G/W DISCHARGE PRESSURE LEVEL","R4")
:
:-------- UTILITY ROUTINE TO SWAP CURRENT RESULTS INTO SAVED RESULTS ------------
:
PROC SAVORIG
:basic 303 & 304 saved in grps96mc.dat
:SWAP(M303,M123)
:TREES GREEN>0
:SWAP(M304,M124)
:PASTURE LAI
SWAP(M241,M201)
:CUMULATIVE RUN-OFF
SWAP(ETAP,M202)
:TOTAL PASTURE ET
SWAP(M243,M203)
:STREAMFLOW INTEGRATED
SWAP(M244,M204)
:TOTAL STORAGE GAIN
SWAP(M245,M205)
:END OF APRIL SHALLOW STORE
SWAP(STORE1,M206)
:FINAL SHALLOW STORE
SWAP(STORE2,M207)
:DEEP STORAGE
SWAP(ETTREE,M208)
:TOTAL TREE ET
SWAP(M249,M209)
:NET TRANSFER SHALLOW TO DEEP
SWAP(M250,M210)
:FLOW FROM DEEP TO SHALLOW
SWAP(M251,M211)
:NET RECHARGE
SWAP(M252,M212)
:FINAL DEEP DRAINAGE
SWAP(M253,M213)
:THROUGHFLOW
SWAP(M254,M214)
:SURPLUS RECHARGE
SWAP(M255,M215)
:INTEGRATED SEEPAGE VOLUME
SWAP(M259,M219)
:TOTAL ET DEEPROOTED PASTURE
SWAP(M260,M220)
```

```
:runoff less evap
SWAP(M448,M224)
:NEGATIVE THROUGHFLOW
SWAP(M265,M225)
:SALT LOAD DISCHARGED KG/YR
SWAP(M266,M226)
:SALT AFFECTED AREA
SWAP(M447,M227)
:POSITIVE DISCHARGE
SWAP(M268,M228)
:CLAY VOLUME
SWAP(M269,M229)
ENDPROC
:
:-------- ROUTINE TO RECALCULATE POTENTIAL TRANSPIRATION ---
: Prepare transpiration maps if pasture area definitions have changed
:
PROC TRANSPIRE
NAME( M439,"PASTURE MAX. TRANSPIRATION(MM)","R4")
    M439 << .352*M12*M304/2.1*clai
:DEEP PASTURE USES MAPS 435, 437
:
NAME(M435,"DEEP ROOT MAX. TRANSPIRATION(MM)","R4")
  M435 << .352*M12*M305
NAME(M437,"DEEP STORE LIMIT FOR DEEP ROOTS","R4")
: CLAY VOL - (ROOT DEPTH - TOP SOIL)*AREA*POROSITY/STD LEAF AREA * ACTUAL LEAF
AREA
  M437 << CLAYVOL-(2-DEPTH)*AREA*.2/DPLAI*M305
ENDPROC
:
:-------- ROUTINE TO MODEL ONE MONTH -----------------------
:
PROC MONTH RAIN EVAP GROWTH MTHNAME
: ADD RAIN TO TOP LAYER
    ISTORE1 << MAX(STORE1,0)
    ISTORE2 << STORE2
:STEP
    STORE1 << STORE1 + M11*RAIN*FACTOR
: PASTURE ET CANNOT CAUSE STORE TO BECOME LESS THAN -DRY
: COMPUTE POTENTIAL ET DEMAND
    LAP << MAX(0, MIN(STORE1+ DRY, EVAP*FACTOR*GROWTH*M439))
    LDRP << EVAP*1.0*M435
    LTREE << EVAP*FACTOR*1.0*M436
: ANNUAL PASTURE OR CROP IS FINAL
    STORE1 << STORE1 - LAP
    ETAP << ETAP + LAP
: DEEP ROOTED PASTURE SHALLOW LAYER DEMAND
    UDRP << MAX(0, MIN(STORE1+ DRY, LDRP))
    STORE1 << STORE1 - UDRP
    ETDRP << ETDRP + UDRP
: POTENTIAL DEEP ROOTED DEMAND ON DEEP LAYER
    LDRP << (LDRP - UDRP)*.6
: TREE SHALLOW LAYER DEMAND
    UTREE << MAX(0, MIN(STORE1+ DRY, LTREE))
    STORE1 << STORE1 - UTREE
```

```
    ETTREE << ETTREE + UTREE
: POTENTIAL TREE DEMAND ON CLAY LAYER
    LTREE << (LTREE - UTREE)*.6
: ADD ALL SHALLOW TRANSPIRATION IN LAP SO TOTAL CAN BE USED LATER
    LAP << LAP + UDRP + UTREE
:    LAP << LAP + UTREE
NAME( M414 ,"INFILTRATION#")
    M414 << MAX(0,MIN(STORE1,M130))
: PASTURE CAN USE CLAY STORE UP TO DEEP ROOT LIMIT (M437 =STOREC VALUE AT LIMIT)
    UDRP << MAX(0,MIN(LDRP,(STOREC - M437)))
: ALL REMAINING TREE ET DEMAND CAN COME FROM CLAY
    STOREC << STOREC - UDRP - LTREE
:    STOREC << STOREC - LTREE
:    MAXSTORE2 << MAX(MAXSTORE2,-STORE2)
    ETDRP << ETDRP + UDRP
    ETTREE << ETTREE + LTREE
: TOP UP CLAY AT UNSAT RATE IF SHALLOW WATER AVAILABLE
    M414 << MIN(CLAYVOL-STOREC,MAX(0,MIN(STORE1,M130)))
    STOREC << STOREC+M414
    STORE1 << STORE1-M414
: ADD DEEP GROUNDWATER FLOW BALANCE (M442) TO STORE2
    STORE2 << STORE2 + M442*((rain/.531-1/12)*gwscale+1/12)
: POTENTIAL RECHARGE TO DEEP IF SHALLOW WATER AVAILABLE AND CLAY SATURATED
    TEMP2 << IF(STOREC>=CLAYVOL-1e-4,MAX(0,MIN(STORE1,CLAYTRANS)),0)
: POTENTIAL FLOW FROM STORE2 TO CLAY LAYER IF CLAY IS TOO DRY
: STORE DEFICIT IS WANTED, LATERAL INFLOW IS AVAILABLE=THROUGHFLOW+G/W FLOW
BALANCE
    TEMP1 << MIN(MAX(MINCLAY-STOREC,0),:
             MAX((M253+M442)*((rain/.531-1/12)*gwscale+1/12),0))
NAME( M251 ,"CUM. NET RECHARGE #")
    M251 << M251 + TEMP2 - TEMP1
    STORE2 << STORE2 + TEMP2
NAME( M422 ,"SURPLUS DEEP STORE #")
    M422 << MAX(0,STORE2)
    M250 << M250 + M422
: STORE2 SURPLUS GOES INTO STOREC FIRST, THEN EXCESS INTO STORE1
    STOREC << STOREC + M422 + TEMP1
    M421 << MAX(0,STOREC-CLAYVOL)
    STOREC << STOREC - M421
: SAVE STORE1 WHICH WILL HAVE SALT ADDED, FOR CONCENTRATION CALCS
    TEMP1 << MIN(WATERST,STORE1)
: add actual deep g/w discharge TO SHALLOW STORE
    STORE1 << STORE1 + M421 - TEMP2
    STORE2 << min(0,STORE2)
GWFS(OUTVOL=STORE1,INVOL=STORE1,DSPRS=M117,SLOPE=M105,:
    KSAT=K,THICK=DEPTH,VOIDS=POROSITY)
NAME( M413 ,"RUN-OFF #")
    M413 << IF(STORE1>WATERST,STORE1-WATERST,0)
    STORE1 << STORE1 - M413
RENAME(STORE1,"FINAL STORAGE MTHNAME #")
    TEMP2 << MAX(M413-M115*MAX(EFRAC*.7*EVAP*FACTOR*M12-LAP,0),0)
    M260 << M260 + TEMP2
    M241 << M241 + M413
 ENDPROC
:
```

```
:------- ROUTINE TO MODEL ONE YEAR --------------
:
PROC YEAR YR
NAME( M440 ,"INITIAL deep and shallow STORAGE")
    M440 << STORE1 + STORE2 + STOREC
NAME( M241 ,"CUMULATIVE RUN-OFF#")
    M241 << 0
    ETAP << 0
    ETDRP << 0
    ETTREE << 0
NAME( M249 ,"NET TRANSFER FROM SHALLOW STORE TO DEEP#")
    M249 << 0
NAME( M250 ,"disch added to shallow store#")
    M250 << 0
NAME( M260 ,"CUM. RUN-OFF LESS EVAP")
    M260 << 0
NAME( M442 ,"POTENTIAL RECHARGE & DISCHARGE#")
: net inflow - outflow (t/flow) = SURP RECH - NET RECH +  DISCH + (neg t/flow only)
    M442 << M254 - M251 + M447 + m448
:NAME( M442 ,"MONTHLY DISCHARGE#")
:     M442 << M442/12
NAME( M251 ,"CUM. NET RECHARGE#")
    M251 << 0
: lemon PASTURE PROFILE
 MONTH      .049 ; .036 ;  .8  ; SEP;
 MONTH      .033 ; .054 ;  .4  ; OCT;
 MONTH      .013 ; .066 ; .3 ; NOV;
 MONTH      .008 ; .086 ; 0 ; DEC;
 MONTH      .005 ; .091 ; 0 ; JAN;
 MONTH      .007 ; .079 ; 0 ; FEB;
 MONTH      .012 ; .070 ;  .2 ; MAR;
 MONTH      .028 ; .041 ;  .5 ; APR;
NAME( M245 ,"END OF APRIL SHALLOW STORAGE#")
    M245 << STORE1
 MONTH      .075 ; .029 ; 1. ; MAY;
 MONTH      .112 ; .022 ; 1. ; JUN;
 MONTH      .106 ; .023 ; 1. ; JUL;
 MONTH      .083 ; .027 ; 1. ; AUG;
:For cells in lakes, & major streams remove annual evaporation
    M260 << M260-AREA/1000*.7*M12*(1-M115)
:sum run-off over catchment and print to file SF.yyr
NAME( M243 ,"STREAMFLOW #")
INTDN(INTMAP=M243,VALMAP=M260,DRAIN=M116)
NAME( M244 ,"STORAGE GAIN #")
    M244 << STORE1 + STORE2 + STOREC - M440
NAME( M[700+YR] ,"STORE1")
M[700+YR] << STORE1
NAME( M[710+YR] ,"STOREC")
M[710+YR] << STOREC
NAME( M[720+YR] ,"STORE2")
M[720+YR] << STORE2
: THICKNESS=TOPLVL-BOTLVL. COMPENSATE FOR THIN TRAMSMISSIVE LAYER BY FACTOR
ON KSAT
 GWFD(DISCH=M252,THRFLW=M253,SURPRCH=M254,NETRCH=M251,:
      DSPRS=M117,SLOPE=M105,HTRANS=KD,BOTLVL=M544,:
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      TOPLVL=M542,VPERM=VC,OUTHD=m269,hdfact=.5)
     M447 << IF(M252>0,M252,0)
     m448 << if(m253<0,m253,0)
name(m262,"shallow water balance","r4")
 m262 << m11*.531-etap-etdrp-ettree-m241-m244+m442
name(m263,"deep flow balance","r4")
 m263 << m251 - m447 - m254 -m448
: deep groundwater balance
: 0 = dd - nr + sr + tf
OVROU(OUTFILE="NR.yyr",MAPA=BNDY,MAPB=M251)
OVROU(OUTFILE="sr.yyr",MAPA=BNDY,MAPB=M254)
OVROU(OUTFILE="DD.yyr",MAPA=BNDY,MAPB=M447)
OVROU(OUTFILE="tf.yyr",MAPA=BNDY,MAPB=M448)
: TOTAL groundwater balance
: rain*0.85*0.625 = rain*.531 = pe + te + ro + SG - gf
OVROU(OUTFILE="PE.yyr",MAPA=BNDY,MAPB=ETAP)
OVROU(OUTFILE="DE.yyr",MAPA=BNDY,MAPB=ETDRP)
OVROU(OUTFILE="TE.yyr",MAPA=BNDY,MAPB=ETTREE)
OVROU(OUTFILE="SG.yyr",MAPA=BNDY,MAPB=M244)
OVROU(OUTFILE="ro.yyr",MAPA=BNDY,MAPB=M241)
OVROU(OUTFILE="GF.yyr",MAPA=BNDY,MAPB=M442)
: other (ad = deep groundwater added to shallow)
OVROU(OUTFILE="SS.yyr",MAPA=BNDY,MAPB=STORE1)
OVROU(OUTFILE="DS.yyr",MAPA=BNDY,MAPB=STORE2)
OVROU(OUTFILE="CS.yyr",MAPA=BNDY,MAPB=STOREC)
OVROU(OUTFILE="SF.yyr",MAPA=BNDY,MAPB=M260)
ENDPROC
:
:------------- SMOOTH OUTPUT MAPS AND CALCULATE SALT AFFECTED AREA ---------
:
PROC SMDISCH
NAME( M422,"SUMMED ADJACENT","R4")
NAME( M421,"CELL COUNT","R4")
NAME( M255,"SEEPAGE VOLUME integrated #","R4")
 INTDN(INTMAP=M255,DRAIN=M116,VALMAP=M447)
 OVROU(MAPA=M150,MAPB=M125,MAPC=M447,OUTFILE="PS.SM")
NAME( M265,"SALT DISCHARGE (KG/YR) #","R4")
     M265 << IF(M125=0,m447*DGWSAL/1000,0)
NAME( M266,"SALT AFFECTED LAND #","R4")
     M266 << IF(M447>0,1,0)
OVROU(OUTFILE="SA.SM",MAPA=BNDY,MAPB=M266)
ENDPROC
:
:-----  PRELIMINARY CALCS WHEN STARTING FROM BEGINNING --------------
:
PROC START
NAME M115 I2 LAKE=0, OTHER=1
     M115 << IF(M25>0,0,1)
NAME( M305 ,"DEEP PASTURE LAI","R4")
TRANSPIRE
NAME( M436,"ANNUAL TREE TRANSPIRATION (MM)","R4")
:            NET RAIN / NATURAL GREENNESS       * ACTUAL GREENNESS
     M436 << 1.4 *.85*M11 / (17.3 + 0.0104*M11) *   M303
:NAME( STORE1 ,"INITIAL STORAGE ")
     STORE1 << WATERST
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:NAME( STORE2 ,"DEEP STORE")
    STORE2 << 0
    STOREC << CLAYVOL
NAME( M442 ,"MONTHLY DISCHARGE #")
NAME( M251 ,"CUM. NET RECHARGE#")
NAME( M252,"UNSMOOTHED DEEP DRAINAGE #","R4")
NAME( M253,"DEEP GROUNDWATER THROUGHFLOW #","R4")
NAME( M254,"SURPLUS RECHARGE #","R4")
: ZERO COMPONENTS OF DEEP GROUNDWATER BALANCE FOR INITIAL YEAR
     M253 << 0
     M254 << 0
     M251 << 0
     M442 << 0
     M447 << 0
     M448 << 0
:MAXSTORE2 << 0
ENDPROC
:----- MAIN PROGRAM ---------------------------
:
: RUN THREE YEARS STARTING WITH FULL SATURATION AT 1ST SEPTEMBER
:SET CLAY LAYERS
NAME(MAP=M542,TITLE="TOP CLAY LEVEL",TYPE="R4")
NAME(MAP=M544,TITLE="BOTTOM CLAY LEVEL",TYPE="R4")
M542 << M21-1.5
M544 << M542-15.5
NAME(M303,"GREEN>0","R4")
NAME(M304,"PASTURE LAI=CLAI","R4")
M303 << M123
M304 << M124
START
m305 << 0
YEAR 1;
YEAR 2;
TEMP1 << (WATERST-M701)^2/(WATERST-2*M701+M702)
STORE1 << IF(TEMP1>0 & TEMP1<WATERST, WATERST-TEMP1, STORE1)
TEMP1 << (CLAYVOL-M711)^2/(CLAYVOL-2*M711+M712)
STOREC << IF(TEMP1>0 & TEMP1<CLAYVOL,CLAYVOL-TEMP1, STOREC)
TEMP1 << M721^2/(M722-2*M721)
STORE2 << IF(TEMP1>0,-TEMP1,STORE2)
YEAR 3;
SMDISCH
SAVORIG
END
```