

Centre for Groundwater Studies

GROUNDWATER FLOW MODELLING TO ASSIST LAND MANAGEMENT IN THE COOKE PLAINS AREA OF SOUTH AUSTRALIA

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Centre for Groundwater Studies Report No. 58
ISBN No. 1 875753 10 9

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1. INTRODUCTION

Substantial increases in land and stream salinity in Australia have resulted from the extensive clearing of native, perennial, deep rooted vegetation for the development of annual shallow rooted agricultural crops and pastures. Salinisation is now recognised as an increasing problem in the dryland agricultural regions of Australia (Williamson, 1990; MDBC, 1992). The change in land use has led to increased groundwater recharge and rising watertables. Salts stored in the unsaturated soil profile have been mobilised and transported to the soil surface where salinisation results.

Dryland salinisation is a growing concern for agriculture in the Coastal Plain of the Murray Basin. As with other areas of the South Australian and Victorian Mallee, several studies have shown that the clearing of native vegetation and replacement with shallow rooted crop and pasture species has led to significant increases in recharge rates (Walker *et al.*, 1992a; Kennett-Smith *et al.*, 1994). The low topographic relief, coupled with shallow, saline groundwaters minimise the lag between the effect of increased recharge rates and watertable response. Watertables have been rising at a rate of 0.05 to 0.1 metre per year as a result of increased groundwater recharge derived locally, as well as from upgradient areas. Salinisation has been occurring within the lowlands over the last decade (Barnett, 1992).

Within the Upper South-East region (which includes the Coastal Plain and the town of Cooke Plains) it is estimated that up to 200,000 ha of land presently shows signs of, or is at future risk of salinisation, with an associated annual cost in production of at least \$7M (MDBC, 1992; Case Study 1).

If the spread of dryland salinity is to be contained remedial measures will be needed. The objective of these measures would be to maintain or lower watertables to depths where they do not reduce agricultural productivity. It will be shown later in this report that only on-site options (ie. within the farm and undertaken by landholders) are feasible in the Coastal Plain and these include recharge reduction (which treats the cause) or discharge enhancement (which treats the effect).

Recharge reduction involves the establishment of agronomic species (usually deep rooted and perennial) which utilise a greater proportion of rainfall (stored as soil water) and thereby limit deep drainage which subsequently recharges groundwater. A well documented example of the success of recharge reduction is the small catchment at Burke's Flat in Victoria, where the establishment of lucerne and other high water use species have resulted in steadily declining watertables (MDBC, 1992; Case study 2). Recharge rates beneath lucerne at Cooke Plains have been found to be comparable to native vegetation (Walker *et al.*, 1992b) and lucerne shows promise as a means of reducing recharge in the study area.

Discharge enhancement involves pumping groundwaters (using windmills) to so called "sacrificial" areas in order to lower watertables in the vicinity of the windmill. This technique has been used successfully in the fractured rock (semi-confined) aquifers of the wheatbelt in Western Australia (Salama *et al.*, 1993). Discharge enhancement was feasible in Western Australia because of the favourable aquifer characteristics (low transmissivity and storage coefficient) and also because only localised areas were salinised (usually upgradient of dykes).

As reported later the high transmissivity, unconfined aquifer (with high storage coefficient) at Cooke Plains would appear to provide only limited opportunity for lowering watertables through pumping.

The success of field-based trials involving alternative land management options will ultimately be determined by long-term monitoring in those catchments. In order to predict the effectiveness of management strategies to control dryland salinity development in a catchment groundwater models are commonly used.

Dryland salinity studies in Australia involving groundwater flow modelling have typically been performed in well-defined catchments in order to evaluate land management options. This often includes evaluating what reduction in recharge rates is required to lower watertables and thus prevent evaporative groundwater discharge and therefore reverse salinisation trends (eg. Morris and Thomson, [1983]; Hookey and Loh, [1985]; Barnett, [1989]; Salama *et al.*, [1990]; Schofield, [1990]; Doherty and Stallman, [1992]; Richardson *et al.*, [1992]; Salama *et al.*, [1993]).

For the wheatbelt of Western Australia Salama *et al.*, (1993a) evaluated, from modelling, the effect of recharge reduction together with discharge enhancement in the first-order Cuballing catchment following extensive field studies (Salama *et al.*, [1992]; Salama *et al.*, [1993b]). Studies of a similar nature were performed in relatively small catchments on the Eyre Peninsula of SA (Richardson *et al.*, 1992) and in the Darling Downs in Queensland (Doherty and Stallman, 1992). Regional groundwater modelling studies such as that of Barnett, (1989) estimated the need for vegetation controls to reduce saline groundwater inflows to the River Murray. Barnett's model included the Coastal Plain area and indicated the potential for severe salinity problems in the long term.

Remedial measures to tackle dryland salinity in the Coastal Plain have not yet been initiated due to the recent advent of the problem and until now the hydrogeological processes have been poorly understood. The costs and benefits as well as the uncertainties associated with their implementation require an evaluation of the likely impacts for a range of options in order to develop cost-effective management plans. Landholders at Cooke Plains require an assessment of the likelihood of success of local (farm-scale) treatments. A modelling approach which provides a finer spatial resolution than a regional model, and has the sensitivity to simulate small-scale processes (such as recharge and evaporation in a salinised swale) is required.

The aim of this study is to develop (by calibrating against field data) a groundwater flow model in order to predict the impact on groundwater levels for land management options which are available to landholders at Cooke Plains for the control of dryland salinity. It considers options which include continuing current land management practices, reducing groundwater recharge and enhanced groundwater discharge.

2. PHYSIOGRAPHIC SETTING

2.1 The Study Area

The Cooke Plains area is within the Coastal Plain, in the south-western margin of the Murray Basin in South Australia. The study area covers 105 km² between Cooke Plains and Coomandook (Fig. 1). Within the study area two experimental sites (referred to as sites 1 and 2) were established (Figs. 2,3).

The present landscape is derived from the late Pleistocene marine transgression which eroded the former Tertiary sediments to form the low-lying undulating Coastal Plain (Barnett, 1992). A cross section which illustrates the hydrogeological consequences of the marine transgression and includes the extent of the study area is shown in Figure 4. The landscape is low and undulating and characterised by stranded and reworked beach dunes with alternating swales. The absence of surface drainage features is another prominent characteristic of the area. The topographic relief is generally less than 10 metres AHD, although dunes may extend to greater than 30 metre. Virtually all of the former mallee native vegetation was cleared for agriculture around 40 to 70 years ago. The dominant landuse consists of the crop-pasture rotation (barley and wheat crops) with no fallow and extensive sheep grazing on improved pastures. Lucerne pastures were common in the area up until the aphid infestation of the 1970's. Thereafter pastures have consisted of shallow-rooted annual species. Approximately three to four percent of the study area is visibly salt affected.

The Cooke Plains area has a semi-arid climate, with cool wet winters and extended hot dry summers. The mean annual rainfall at Cooke Plains is 380 mm, and pan evaporation is around 1,700 mm yr⁻¹. High variability in rainfall is characteristic of semi-arid areas, and at Cooke Plains the annual rainfall has ranged from 184 mm to 598 mm between 1967 and 1992 (Fig. 5).

2.2 Soils

The soils in the Cooke Plains area have been mapped in earlier regional surveys by Taylor and Poole, (1931) and de Mooy, (1959) and in more detail at site 1 by Hollingsworth *et al.*, (1993) and at site 2 by Dutkiewicz, (1992). A generalised soils association map adapted from de Mooy, (1959) is shown in Figure 6. In the study area, soils are relatively homogeneous due to the dominant presence of the Coolinong Association. The Coolinong Association is characterised by deep sandy profiles, often with a clayey subsoil horizon. The sandy surface soils have been observed to display water repellency. Calcrete, ranging in form from small nodules to thick blankets is present at the surface or at shallow depth in most areas. Within the discharge areas the soils are saline and consist of swampy sands, muds or clays.

2.3 Discharge Areas

Evaporative groundwater discharge features have evolved in the low-lying lands of the Cooke Plains embayment (von der Borsch and Altmann, 1979). These occur in two distinct forms, the

Figure 1: The Cooke Plains - Coomandook area including model area and experimental sites

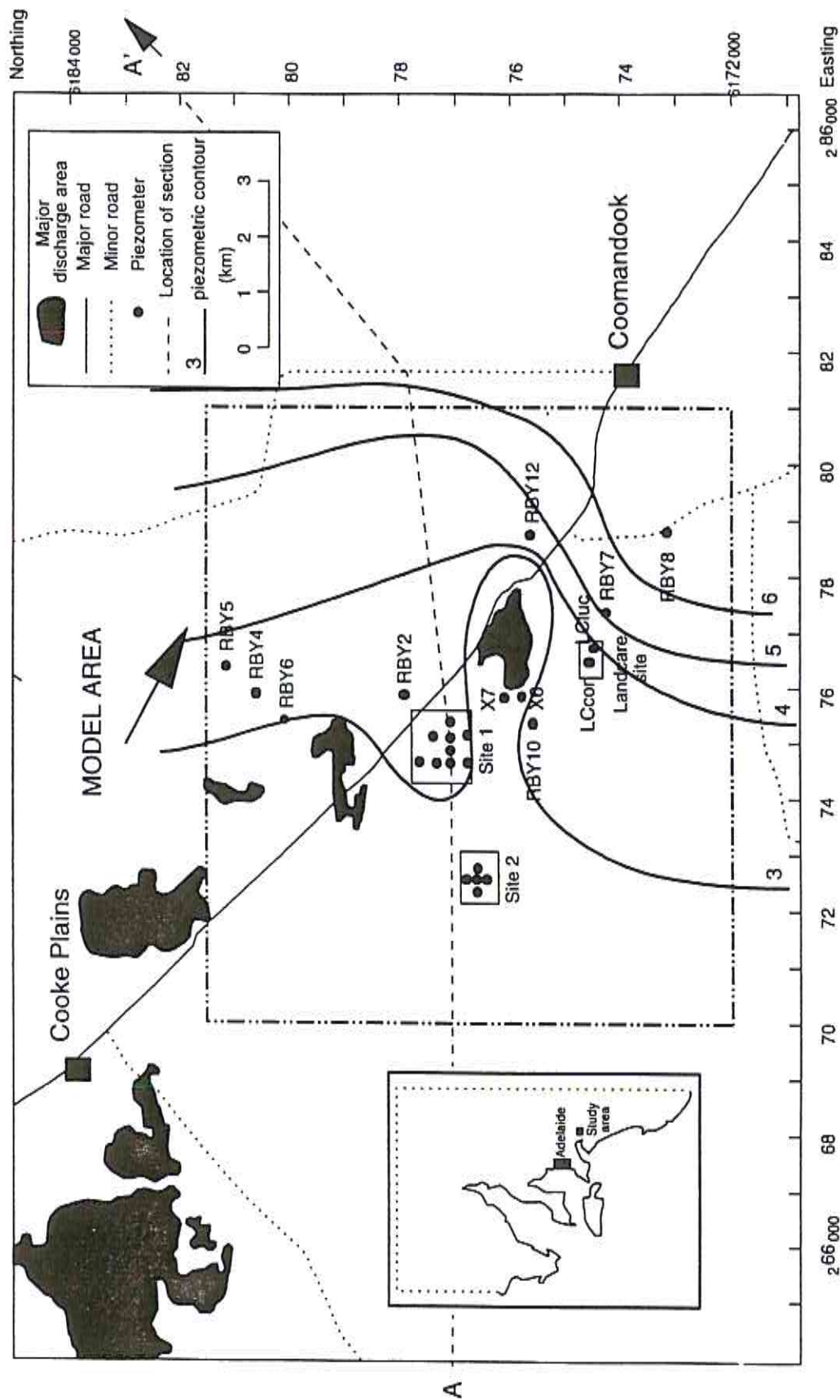


Figure 2: Site plan for experimental site 1 including topographic relief and piezometers

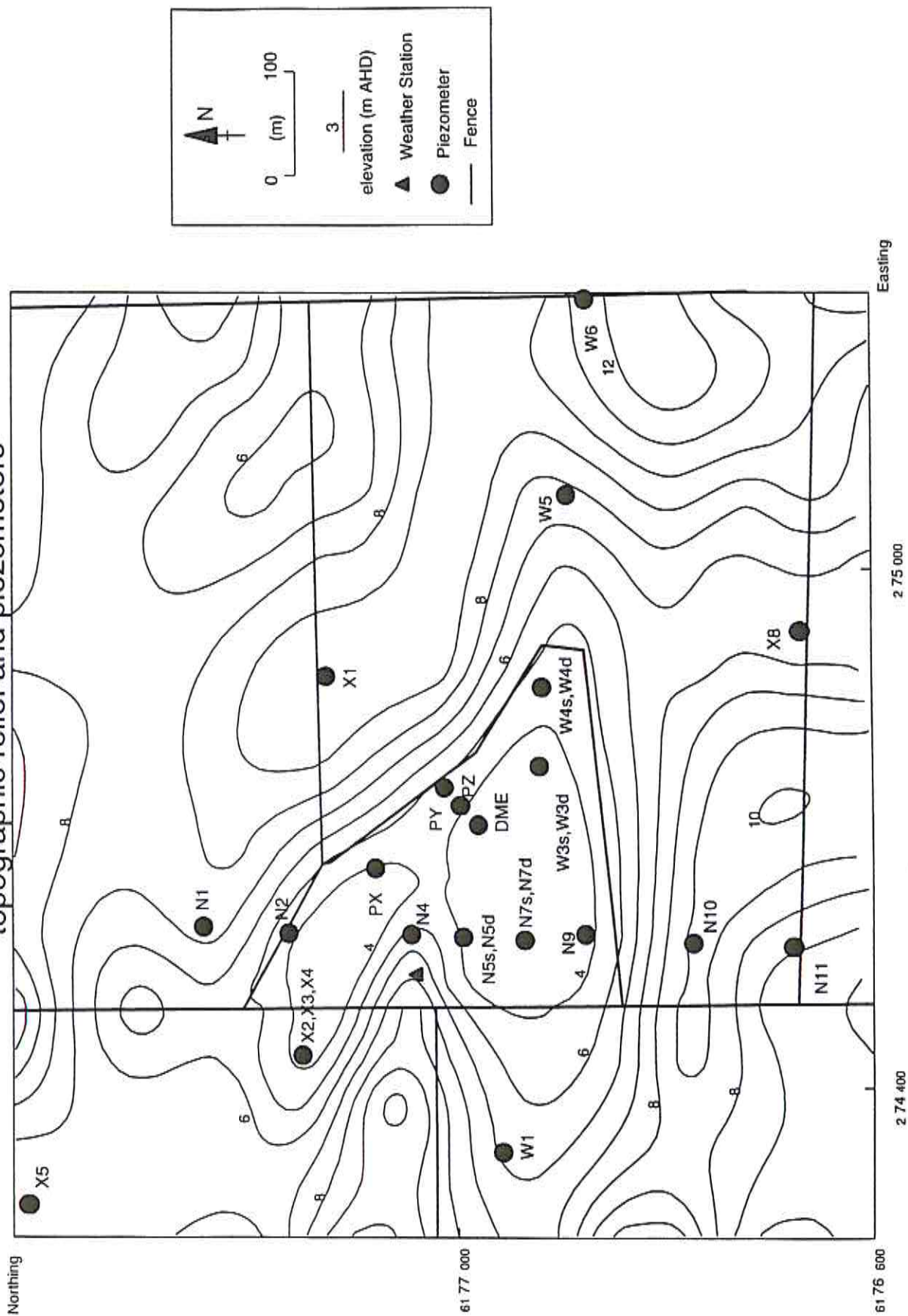


Figure 3: Site plan for experimental site 2 including topographic relief and piezometers

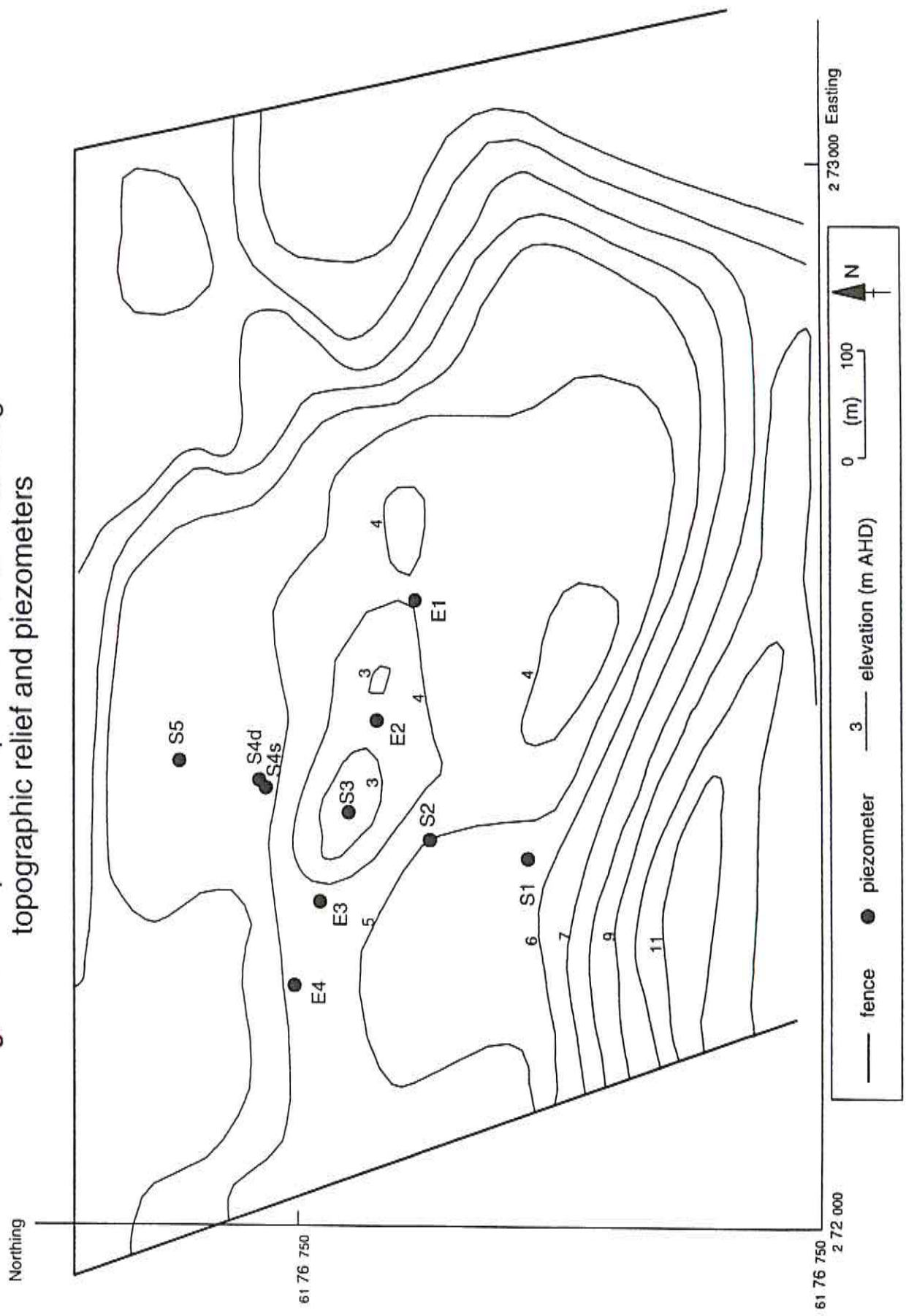


Figure 4: Regional Hydrogeological Transect

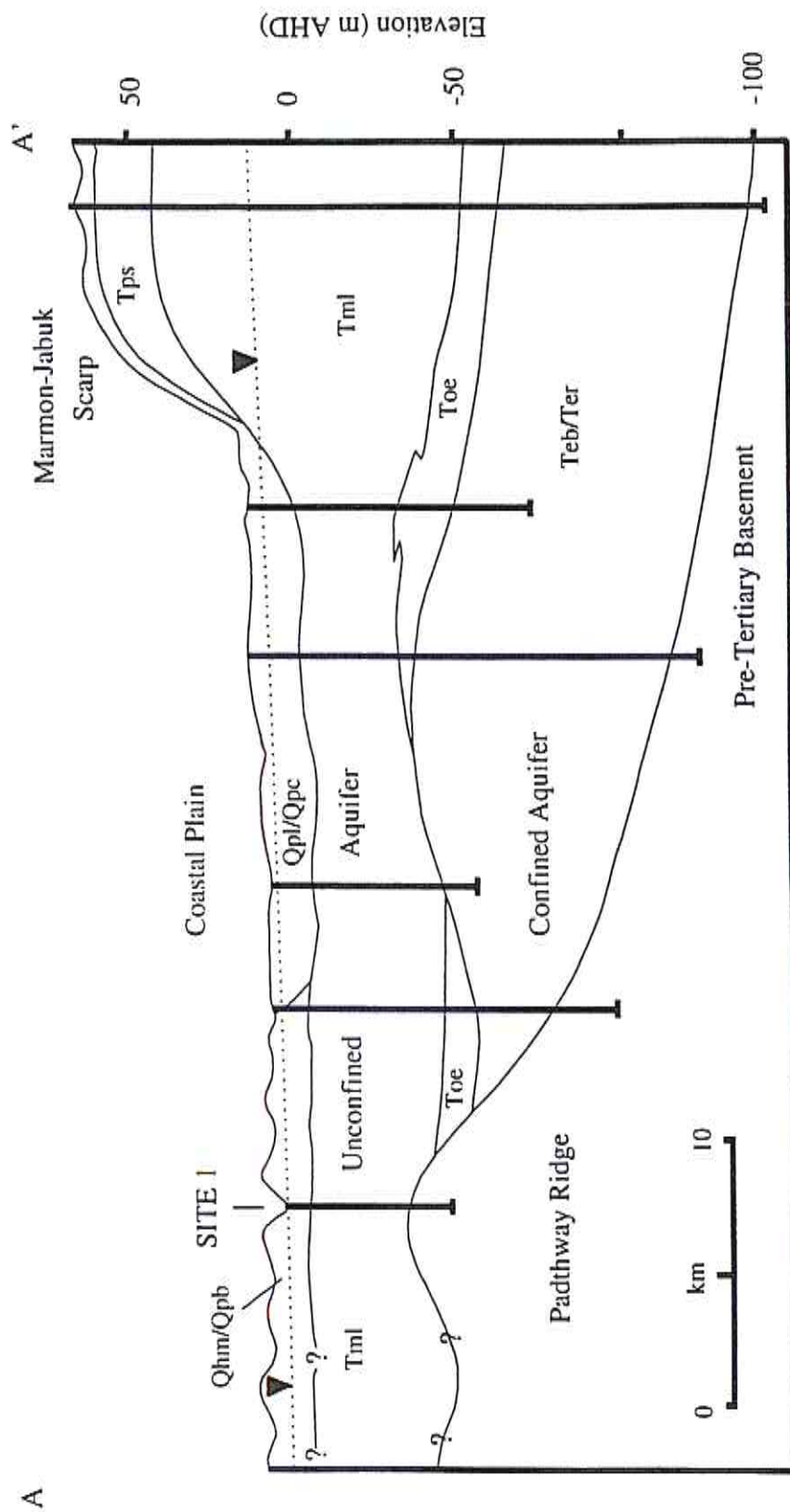


Figure 5: Annual rainfall distribution
at Cooke Plains (1967-1992)

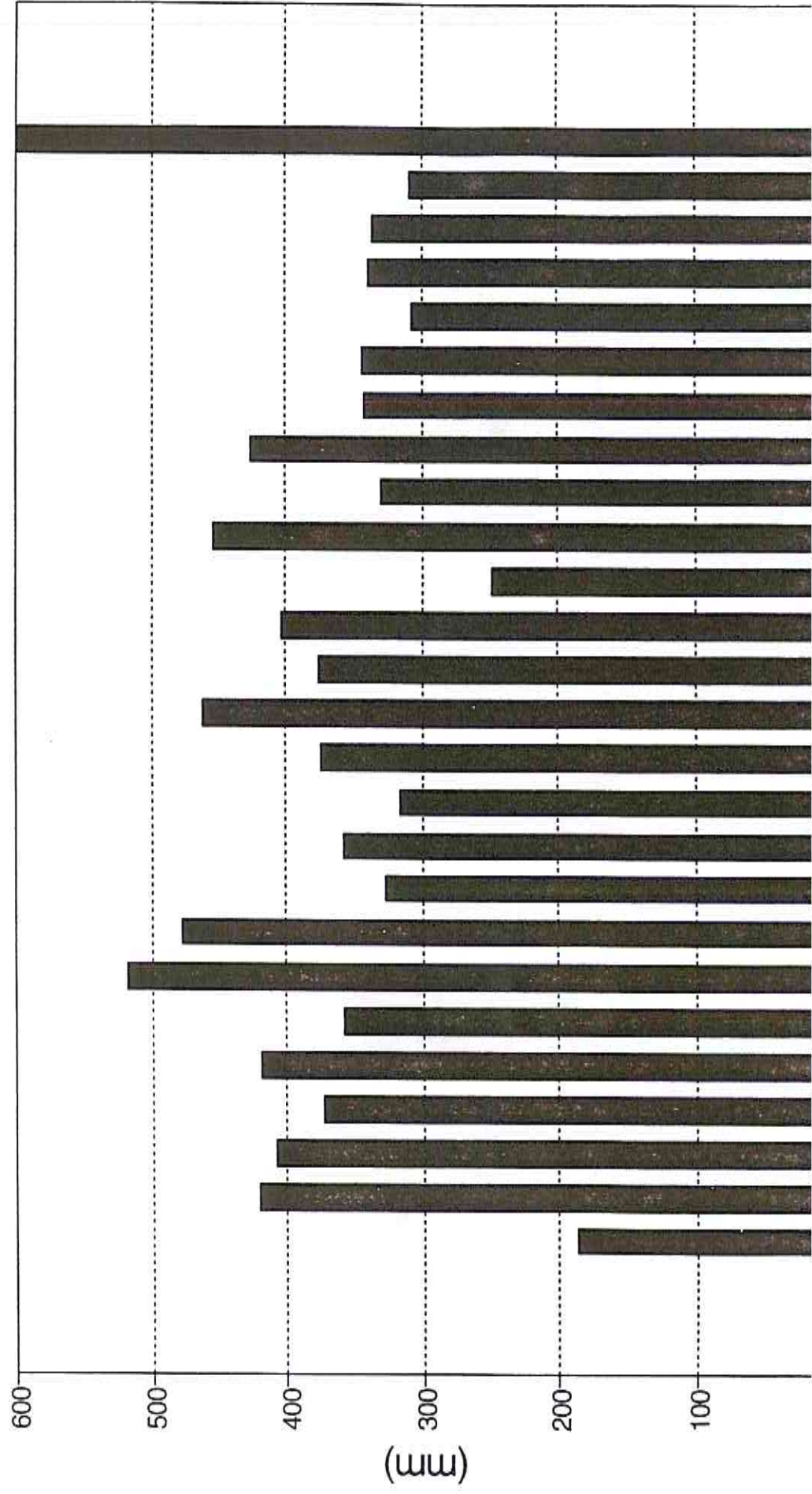
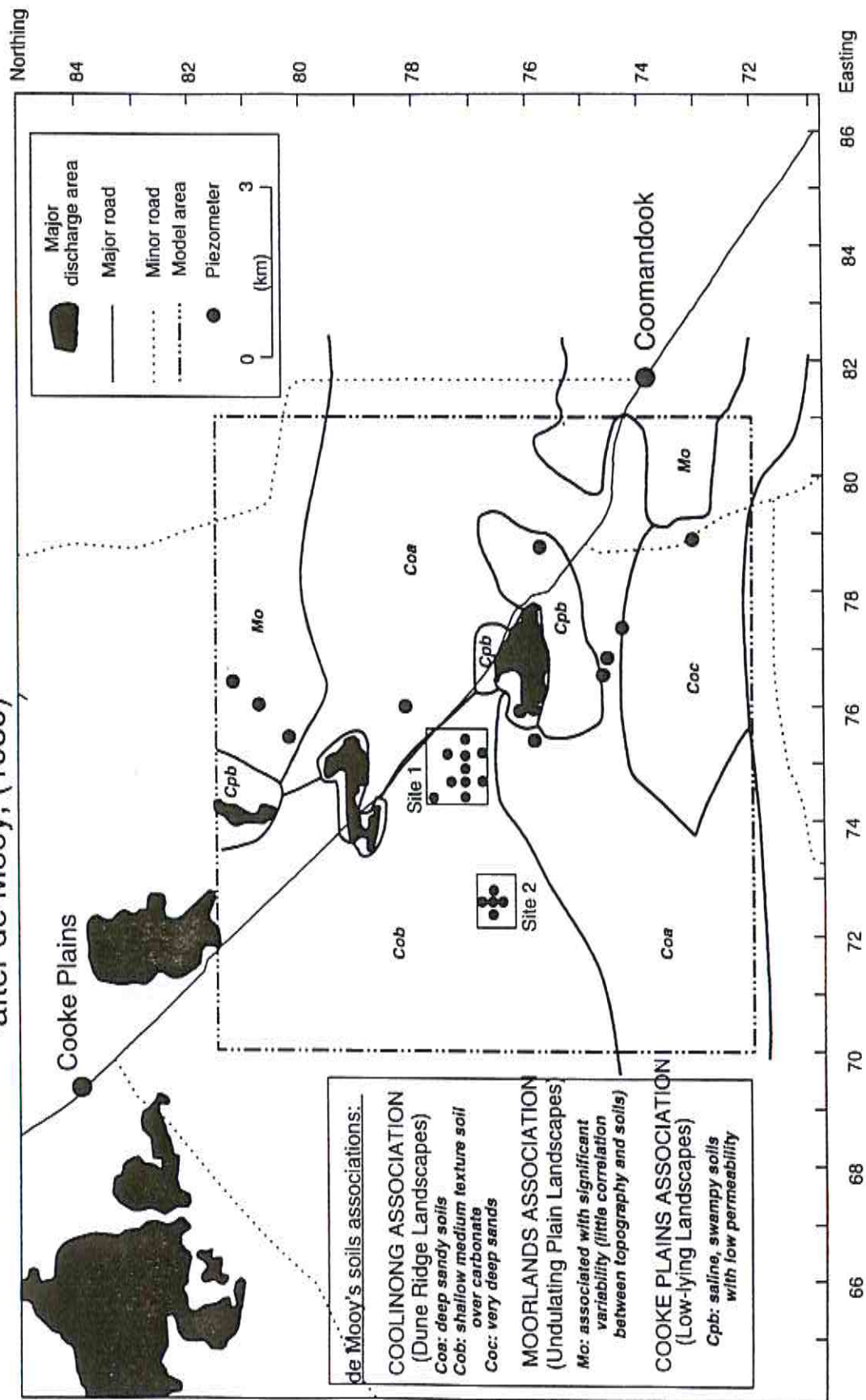


Figure 6: Soil associations of the Cooke Plains - Coomandook area
after de Mooy, (1959)



older, pre-European (and generally larger) discharge areas occur within extensive low-lying flats. The more recent (smaller) discharge areas are a result of secondary salinisation, and occur within the swales. The impact of discharge areas on the groundwater system is discussed later.

2.4 Hydrogeology And Groundwater Flow

The regional hydrogeology is described in detail by Barnett, (1992). The north-south trending Padthaway Ridge forms a basement high and occurs at an elevation of -38 m AHD at site 1 (the only site drilled to basement in the study area), with the deeper Buccleuch (Teb) and Renmark Group (Ter) systems not present in the study area (see Fig. 4).

The regionally extending Murray Group (Tml) aquifer is approximately 20 metres thick and consists of limestone. The Murray Group and overlying Bridgewater (Qpb) and Coomandook (Qpc) Formations are geologically similar and apparently hydraulically interconnected (Fig. 4). The watertable may lie in either of these two units, and together they comprise the unconfined aquifer system.

The Cooke Plains area is a groundwater discharge zone for the regional flow system which originates in the Wimmera, 200 km to the south-east. Groundwater movement is from the east towards Lakes Alexandrina and Albert, under very low hydraulic gradients (Fig. 1). The groundwaters are typically saline (10,000 to 50,000 mg L⁻¹ TDS). Higher salinity waters are associated with evaporative enrichment in the low-lying (discharge) areas, whilst the elevated dunes are associated with relatively less saline waters which occur due the absence of groundwater discharge and enhanced recharge in areas with lighter textured soils.

In areas where the confined aquifers of the Renmark Group are present upward leakage to the unconfined aquifer is low (less than 1 mm yr⁻¹, Barnett 1992) due the low saturated hydraulic conductivity of the confining beds.

3. HYDROLOGIC PARAMETERS

The successful development of a groundwater flow model requires firstly, a conceptual understanding of the dynamics of the groundwater system and secondly, a considerable amount of field data to define aquifer properties, boundary conditions and fluxes, and piezometric data to enable calibration.

3.1 Piezometric Data

The experimental sites and the SA Department of Mines and Energy (DME) regional groundwater monitoring network comprise all of the available piezometric data, which totals 49 piezometers. Nine DME piezometers (established in 1987) occur within the study area. Piezometers were installed at two experimental sites (1 and 2) in 1990 and 1991 respectively and the Coomandook Landcare site was also established in 1990. The location of the sites and piezometers are shown in Figures 1-3. The piezometric data is provided in Appendix 1.

3.2 Aquifer Hydraulic Properties

The aquifer hydraulic properties are based on a 72 hour aquifer pump test undertaken by Barnett, (1992) at site 1. The transmissivity is estimated to be in the order of $300 \text{ m}^2 \text{ day}^{-1}$, and with an aquifer thickness of 20 metres, would suggest the hydraulic conductivity to be around 15 m day^{-1} . The specific yield is approximately 0.05. The base of the aquifer was considered horizontal (-15 m AHD). The groundwater system was modelled as a single unconfined aquifer.

Generally, only a single value of hydraulic conductivity and specific yield were used throughout the model due to the lack of alternative data. At later stages of modelling, the sensitivity to hydraulic conductivity and hence transmissivity was tested.

3.3 Groundwater Recharge

Recharge studies were undertaken using soil water and chloride balances in the unsaturated zone, and were summarised by Walker *et al.*, (1992b). Field investigations involved coring beneath different landuses including crop-pasture rotation, perennial pasture, lucerne, tagasaste, lupins and mallee. The method determines the mean annual recharge rates, which were found to range from 1 to 40 mm yr^{-1} . Recharge was found to be very dependent on landuse, and less dependent on soil texture (presumably because of the similarity in soil type in the area). Recharge rates at the cored sites are provided in Table 1. The average post-cleared recharge rate beneath crop/pasture is considered to be about 20 mm yr^{-1} .

A hydrograph separation technique was used to describe the spatial and temporal variability of recharge.

Table 1: Mean annual recharge rates beneath different vegetation species

LANDUSE	SOIL TYPE	SOILS ASSOCIATION	NUMBER OF MEASURED SITES	MEAN ANNUAL RECHARGE (mm)
CROP/PASTURE	* sands	Coa, Cob	6	5-40
PERENNIAL PASTUR	sands	Coc	1	13
LUCERNE	sands	Coa	2	3
MALLEE	sands	Coc	2	1
* often sandy clay or calcrete horizon present				

Temporal estimation

The piezometric rise during the year is positively correlated with winter rainfall, (r_w) for April to September, as indicated by the five piezometers plotted on Figure 7. The regression of piezometer RBY2 provided the best correlation ($R^2=0.79$), and as it was a reasonable approximation of the other data, was used to estimate the annual recharge rates, (R) in the model. If we assume that the mean annual recharge of 20 mm (from soil coring studies) is equivalent to the product of the mean piezometric rise (from all piezometer data) and specific yield ($s_y=0.05$), then we can determine recharge rates in any year from the equation based on the regression line shown in Figure 7, according to the winter rainfall for that year. The equation used was:

$$R = s_y(0.002r_w - 0.07) \quad (1)$$

Annual variation in watertable rise is significant, and calculated recharge may range from 10 to 40 mm yr⁻¹ between dry and wet years (beneath crop/pasture systems). Recharge was calculated quarterly from equation (1), and values for the modelling (expressed as a fraction of the mean annual recharge) are given in Table 2.

Spatial estimation

Spatial variability of recharge within the study area is principally the result of variations in soil type and landuse. The undulating topography (refer to Fig. 22) creates an effect which is not unlike a mosaic of deep drainage water propagating towards the watertable. From the piezometric rise at sites throughout the model (Fig. 8), it is apparent from the piezometric data that for a given year, recharge rates are relatively uniform over most parts of the landscape. In general, low-lying and upland areas would appear to contribute similar fluxes to the watertable. In the absence of detailed knowledge of the spatial variability of specific yield and recharge it was decided that uniform recharge rates (and specific yield) would represent a reasonable approximation for a model at this scale.

Unsaturated zone storage effects are likely to be small as the sandy soils are free draining and the watertable over most of the area is shallow. Storage effects may need to be considered in more elevated areas (say greater than 20 m AHD), however these areas constitute only a small portion of the study area.

3.4 Groundwater Discharge

The piezometric contour maps show that evaporative discharge has a significant impact on groundwater levels and flows at the local and the broad scale (Fig. 1). Net annual discharge rates in the order of 500 mm yr⁻¹ and 100 mm yr⁻¹ were estimated for the older and recently salinised areas respectively. Higher discharge rates for the older features are a result of the greater proportion of open water available for evaporation. Because of their size total evaporation greatly exceeds the evaporation from recently salinised swales.

Figure 7: The relationship between piezometric rise and seasonal rainfall

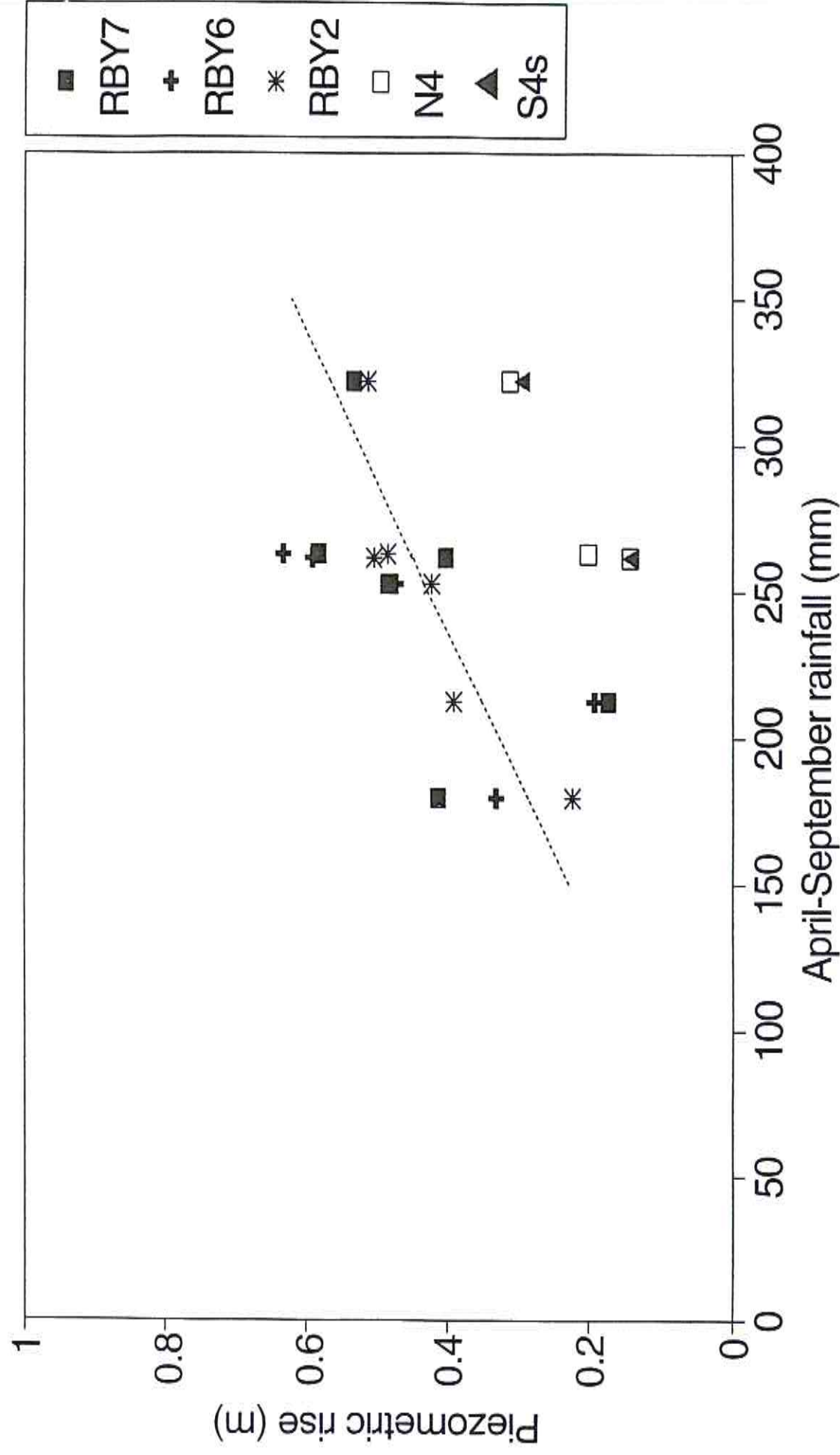
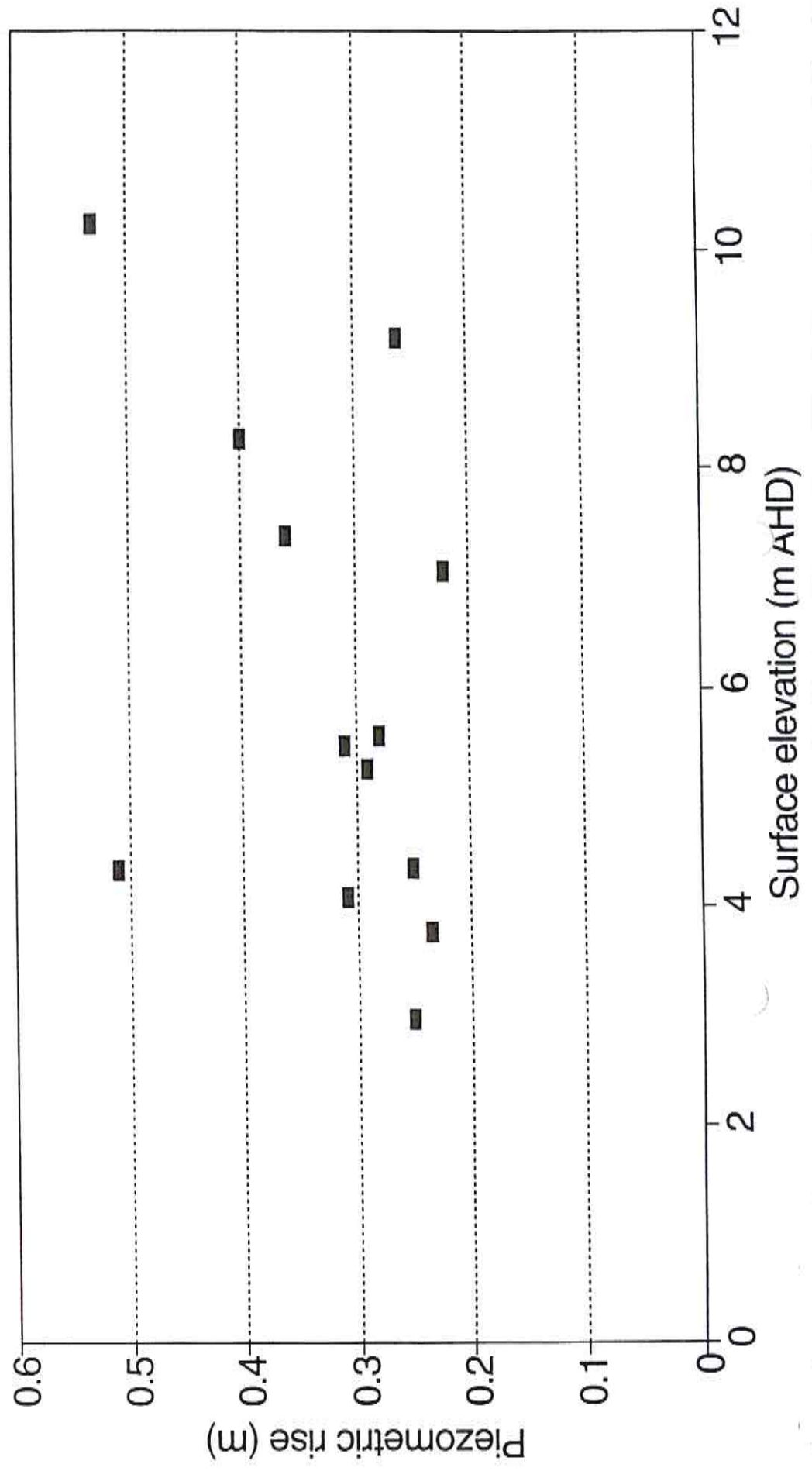


Figure 8: The relationship between piezometric rise and surface elevation



Groundwater discharge is assumed to be described by an evaporation function of the soil. This relates the discharge flux to the depth to watertable below the soil surface. Evaporative groundwater discharge is highest for open water bodies, and is assumed to decrease linearly to the extinction depth in the model.

Electromagnetic investigations (using the shallow penetrating electromagnetic device) in conjunction with soil and crop investigations have identified the groundwater extinction depth. The extinction depth was found to be reasonably shallow in the sandy soils, and is estimated to be in the order of 0.6 to 1.3 metres (Pavelic *et al.*, 1994).

Meteorological data provided the mean potential evaporation rates, which are assumed to be uniform for each year (Table 2).

3.5 Topographic Relief

Detailed topographic surveys were undertaken at sites 1 and 2, and also for the landcare site, and maps drawn at 1:1,250 scale (50 cm contour intervals). Ground surface elevations are also known at sites where piezometers were surveyed. The mean topographic elevation for each cell within the model was estimated, and often extrapolated on the basis of this data as well as 1:50,000 topographic maps (10 m contour intervals) and aerial photographs.

In low-lying areas accurate surface elevations are required to determine the extent and magnitude of groundwater discharge. However in upland areas accurate levels for topographic elevation are not required as discharge does not occur where the watertable is deep, and recharge has been found to be relatively independent of depth to watertable.

Although coverage of areas where detailed surveys were carried out was small, they included most of the areas which are currently susceptible to groundwater discharge. A notable exception is the smaller discharge lake where surface elevations were assumed to be similar to the larger discharge lake. Localised cases of salinisation (say less than one hectare) could not be easily identified from aerial photography, however a reconnaissance survey of the study area revealed few such areas, and therefore they are not considered significant on the broad scale.

4. NUMERICAL MODELLING OF GROUNDWATER FLOW

4.1 Conceptual Model and Boundary Conditions

The conceptual model and groundwater budget which was developed to assist with the modelling is synthesised from the field studies in conjunction with simple analytical models (Appendix 2). A summary of the fluxes of the major components of the groundwater budget is also given.

The area considered in this study consists of a relatively small parcel in comparison to the regional catchment. Conceptually, it may be considered as a small window within a much larger framework. By extending the size of the window we are better able to identify processes which occur within the broader context. In our case, the extent of this window is constrained by logistical, rather than by physical boundaries.

The boundary conditions were based on the piezometric surface maps (see Fig. 1) and hydrograph responses. Contour lines were interpolated from piezometric data from outside of the study area from Barnett, (1992). The model domain was extended beyond the study sites to include an area where boundary conditions could be determined from the piezometric maps. Extending the study domain would minimise the impacts associated with imposing model boundary conditions, which over the time frames considered should not influence the model results (groundwater velocity calculations are given in Appendix 2). The model domain was 9.5 km (N-S) by 11 km (E-W) and included the area from 6172000 to 6181500 N and 270000 to 281000 E (Fig. 1).

The expansion of the model domain beyond the boundaries of the experimental sites would have the effect of making the model more robust, as well as making the study more applicable in the broader context. Whilst the proportion of saline area, aquifer characteristics and groundwater fluxes and boundary conditions may vary in other areas, the general outcomes of the study are likely to be relevant in other areas of the Coastal Plain.

The northern and southern perimeter of the model are zero flux ("no flow") boundaries. The western (downgradient) perimeter is represented as a constant head to allow outflow through the model. At the eastern (upgradient) perimeter a constant head boundary condition and later a general head boundary condition was assigned. The general head boundary is a head dependent flow boundary (Cauchy condition) which allowed heads in the upstream boundary to increase at the current rate of 0.05 m yr^{-1} . These boundary conditions are consistent with the piezometric surface map (Fig. 1) and piezometer hydrographs (Barnett, 1992).

4.2 The MODFLOW Code

The US Geological Survey modular, finite difference groundwater flow model MODFLOW (McDonald and Harbaugh, 1988) was chosen to simulate groundwater flow. Owing to its modular structure it is simple to use, and recharge, evapotranspiration, discharge, wells and drainage together with surface-water interaction can be readily simulated.

4.3 Measurement of Model Performance

In order to ensure the model was performing satisfactorily over the local and broader scale several means for comparison are to be used which include:

- calculation of the RMS error at the local (study site) as well as the larger (throughout model) scales
- plotting piezometer hydrographs and heads at particular times during the simulation
- plotting spatial distribution of heads across local and larger scale transects

The root mean square error (RMS) provides one method of assessing the performance of the model by comparing the computed heads (h_c) with those of the field (measured) heads (h_m). This is given as:

$$RMS = \sqrt{\frac{1}{n} \sum (h_c - h_m)^2} \quad (2)$$

Much of the available piezometric data has been used for the calibration of the model, which extends over a 6 year period from January 1987 to December 1992. Selected piezometric data are presented here only to show particular processes and for comparison with modelled results. Under these circumstances a validation of the model is not possible, although a considerable degree of sensitivity analysis is performed to evaluate the capabilities of the model.

4.4 The Modelling Approach

A two phase modelling approach was adopted. This included an initial "broad-mesh model" (600 cells, with the smallest being 125 m x 125 m) and later this model was superseded by the "fine-mesh model" (2773 cells, with the smallest being 25 m x 25 m). Both models covered the same domain, but differed in their level of discretisation and the method used to deal with evapotranspiration. As a result, the effort put into calibration and sensitivity analysis differed. The amount of effort applied to each model was dependent on the level of detail in that model.

For transient simulations a stress period of 3 months was chosen to allow seasonal recharge and discharge to be simulated. For each stress period there were 20 time steps in which solutions were calculated with a time step multiplier of 1.1 and convergence criterion of 0.001 m. Solutions were extracted only at the end of each stress period.

5. BROAD-MESH MODEL

The broad-mesh model was developed to test our conceptual model of the groundwater system, to test the sensitivity of the model to values of selected parameters, and to select and fine tune the appropriate boundary conditions.

The broad-mesh model had dimensions of 25x24 (600 cells), with grid sizes ranging from 125 metres up to 500 metres. The grid design was such that the finest discretisation was provided around the discharge lakes and experimental sites (Fig. 9). Approximately three percent of the model was specified as salinised and included the older discharge areas (2.8 km²) and salinised swales (0.1 km²). The remainder of the (104.5 km²) model represented recharge areas.

The eastern and western boundary cells were situated approximately along the 6 metre and 2.5 metre contour lines of the piezometric surface (see Fig. 1), and these constant head values were applied to these boundaries. A value of 3.0 metres was applied within the model as starting heads for steady-state runs. Recharge was distributed uniformly across the model, except for the evaporative areas where a net discharge rate was given to simulate evaporative discharge (specified as negative recharge in the model).

Individual model parameter (including recharge, discharge, transmissivity and downgradient boundary values) were adjusted between successive runs to obtain solutions for hydraulic head. The choice of parameter values were often extreme to simulate a wide range of groundwater stresses (particularly for recharge and discharge). Nineteen piezometers across the model domain were selected for the comparison with computed heads (listed with coordinates in Appendix 3 and shown in Figs 1-3). For cases where several piezometers occurred within a particular cell only one piezometer was selected for comparison.

5.1 Steady State Simulations

The steady state simulations were undertaken to provide a first-order sensitivity analysis of the computed heads to the model parameters, and also to provide initial conditions for the subsequent transient simulations. A variety of configurations were tested in the steady-state and transient mode and the results for all runs provided in Appendix 3. Computed heads for the steady-state model were compared with the mean measured head for 1991-92 to calculate the RMS error.

The wide (and often unrealistic) range of values specified in the steady state runs often resulted in large RMS errors (sometimes greater than 1.0 metre), especially where discharge rates were low. A reasonable match between computed and measured heads (RMS=0.38 m) was derived for a recharge rate of 20 mm yr⁻¹, discharge rate of 500 mm yr⁻¹ (older areas), 100 mm yr⁻¹ (salinised swales) and hydraulic conductivity of 20 m day⁻¹ (transmissivity = 340-420 m² day⁻¹). Slightly better RMS errors (down to 0.3 m) were obtained for conditions which were not considered representative of the study area. For example, reducing the recharge rate to 5 mm yr⁻¹ and doubling the hydraulic conductivity provides a RMS of 0.35 m.

An example of the sensitivity of the steady state model to recharge is shown in Figure 10, for

Figure 9: Broad-mesh grid for study area

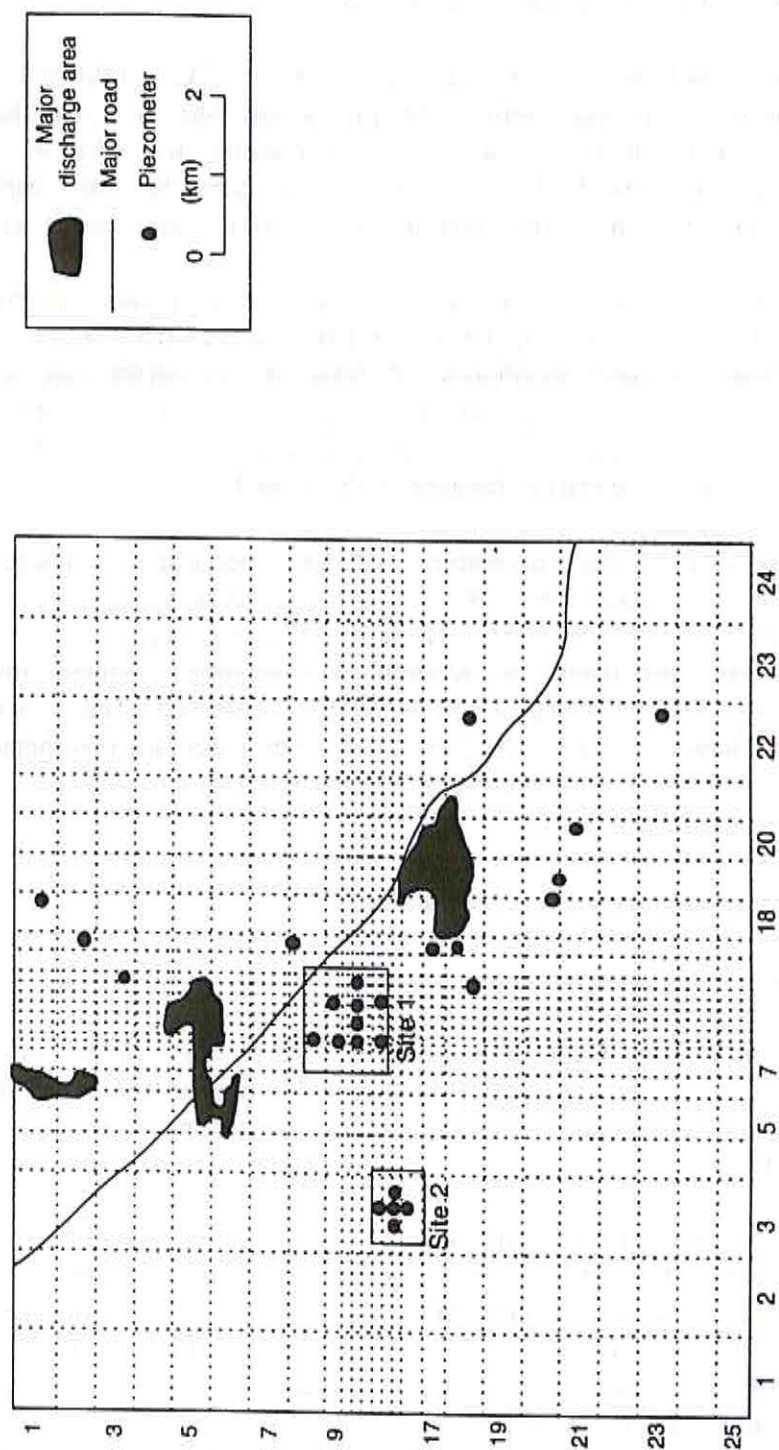
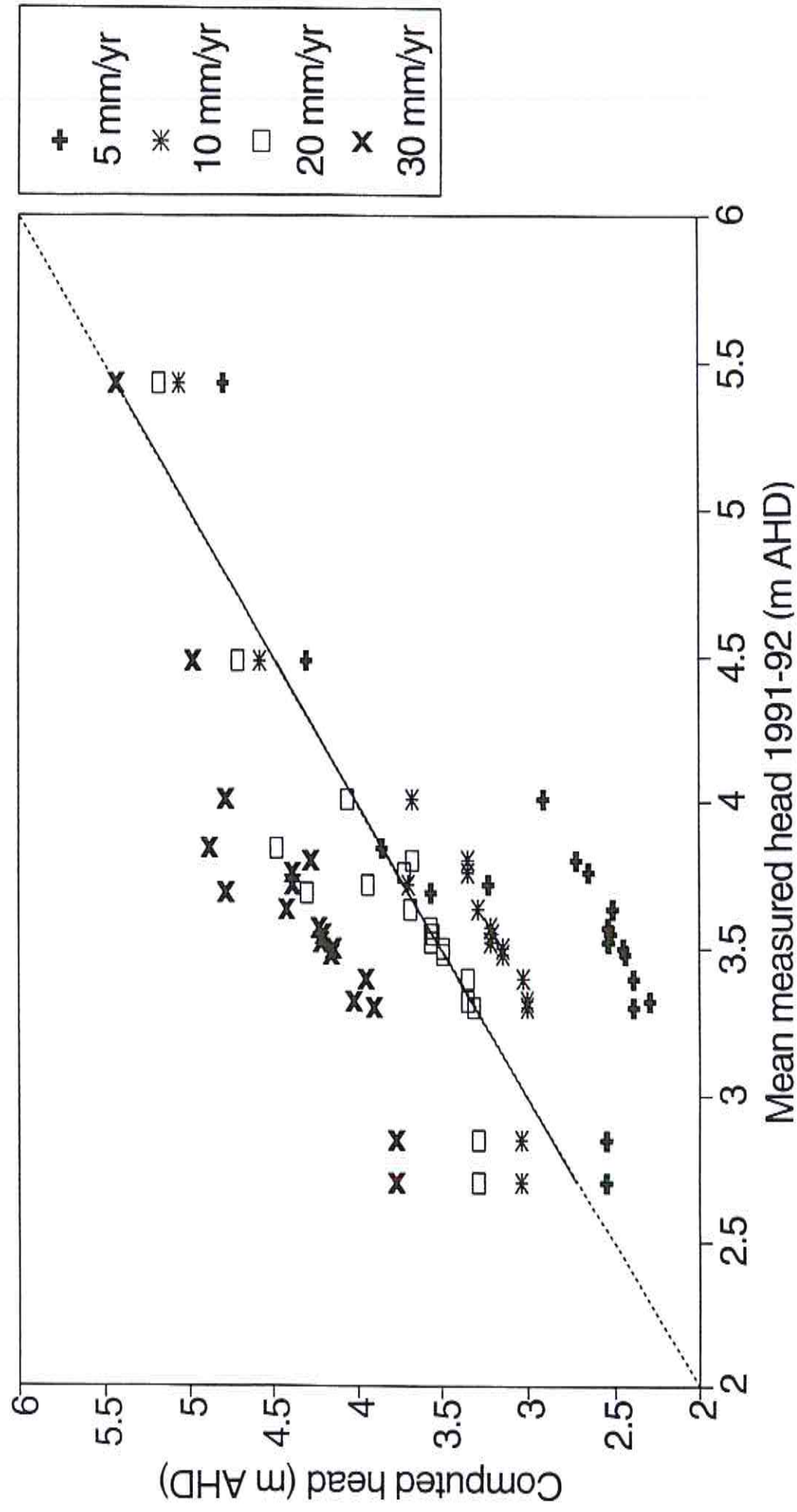


Figure 10: Effect of varying recharge on steady state heads for broad-mesh model



four cases with recharge rates ranging from 5 to 30 mm yr⁻¹. Differences in head of greater than 2 metres can occur over this range of recharge. The steady state simulations had achieved a reasonable match with measured heads. Discrepancies are attributed to the coarse grid size, which was unable to simulate all of the process in sufficient detail.

5.2 Transient Simulations

The model was run in transient mode to simulate the temporal response of the groundwater system to applied stresses. The transient runs used steady state heads as initial conditions. A total of 17 transient simulations were performed, with time scales ranging from 0.5 to 18 years. The initial 6 year period referred to the period from 1987 to 1992, whilst the following 12 years represented a predictive period in the simulation (notionally referred to as years 1993 to 2004). Over the whole period annual recharge and discharge rates considered in the model were generally uniform.

The sensitivity of the model to a number of parameters was tested. In all cases this was done by plotting hydrographs or contour maps to gain a feel for the effects of these parameters.

The sensitivity of the model to recharge was made apparent in the steady state runs, but transient runs identified that very small changes in recharge (of only 5 mm yr⁻¹) can influence the long term trend in the computed hydrographs. Figure 11 shows the measured and computed heads at four sites in the model for recharge rates of 15 and 20 mm yr⁻¹. Whilst seasonal peaks are inherently greater for the higher recharge rate, more importantly the long trend is stationary at a recharge rate of 15 mm yr⁻¹ rising up to 0.05 m yr⁻¹ for a recharge rate of 20 mm yr⁻¹. The fit between measured and computed heads is poor as a result of the approximations made in the model.

A specific yield of 0.05 was found to be the most appropriate value over the range from 0.02 to 0.1 tested by comparing observed and computed annual piezometric level fluctuations, when the recharge rate was 20 mm yr⁻¹.

The sensitivity of the model to recharge and discharge, and their inter-relationship was evaluated in subsequent simulations. In the first case recharge rates for the first 6 years were 20 mm yr⁻¹, and reduced for years 7 to 18 to 2 mm yr⁻¹ (comparable to establishing lucerne throughout most of the model area). As discharge rates are dynamic (ie. vary with changes in watertable depth), best estimates of evaporation rates in discharge areas were specified. Evaporative discharge was steadily reduced from 500 to 200 mm yr⁻¹ in the older discharge areas and from 100 to 40 mm yr⁻¹ in the swales from within 6 years of the decline in recharge. Figure 12 shows a reduction in heads occurring within a few years of the decline in recharge (up to 80 cm in some areas). However the evaporation rates in the older discharge areas decline to a point where groundwater levels recommence rising when the total water flowing out of the model (discharge plus lateral outflow) is less than the total inflow of water out of the model (recharge plus lateral inflow). In reality this would not be the case and watertables would be reduced more rapidly at first and then stabilise as discharge rates adjust in accordance with the new groundwater conditions.

Figure 11: Computed and measured heads
for recharge rates of 15 and 20 mm/yr

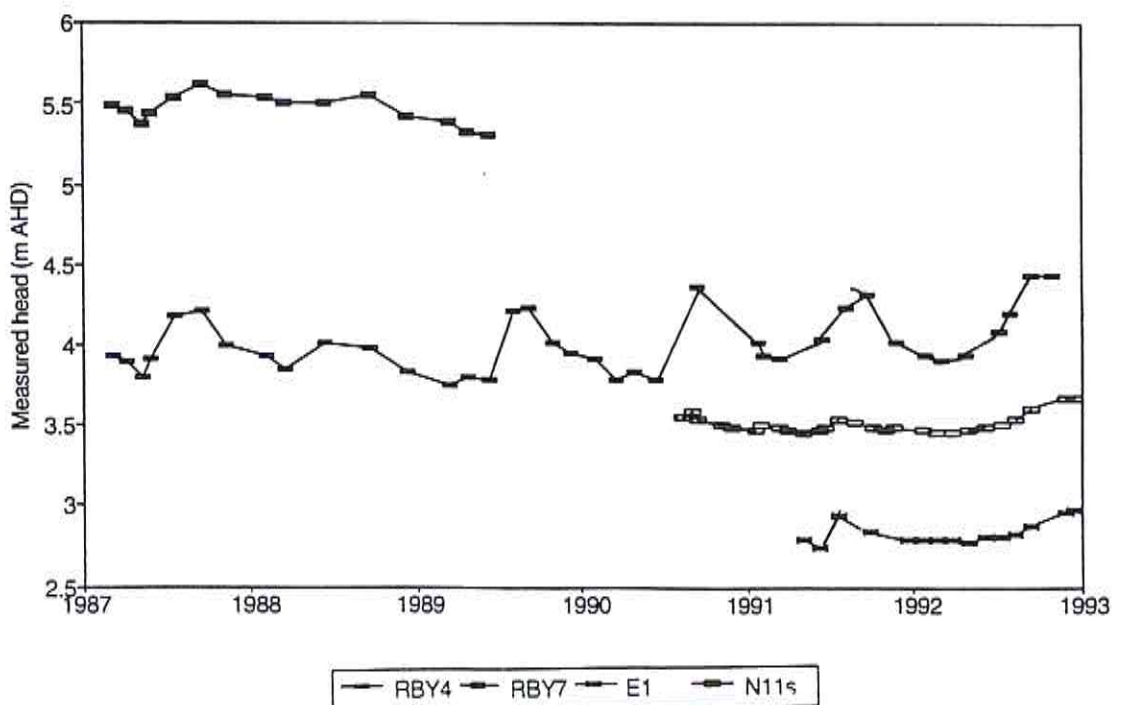
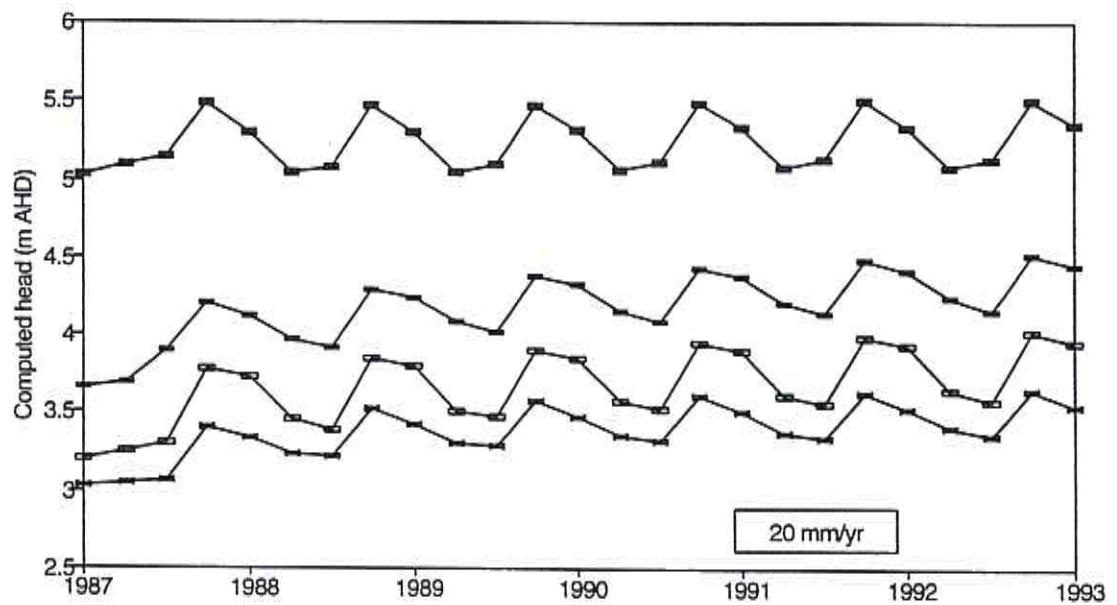
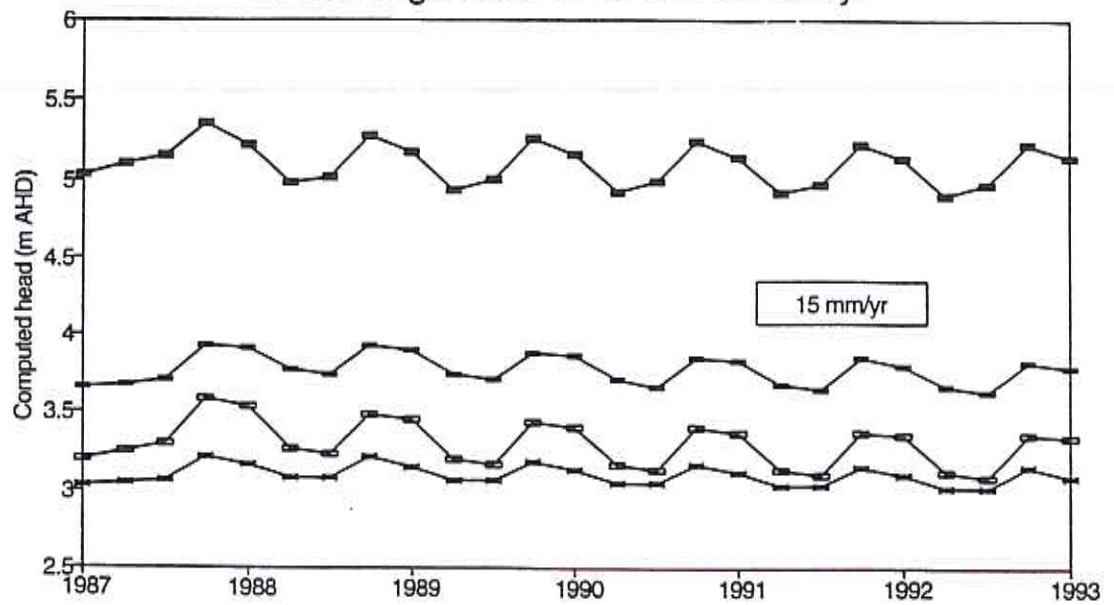
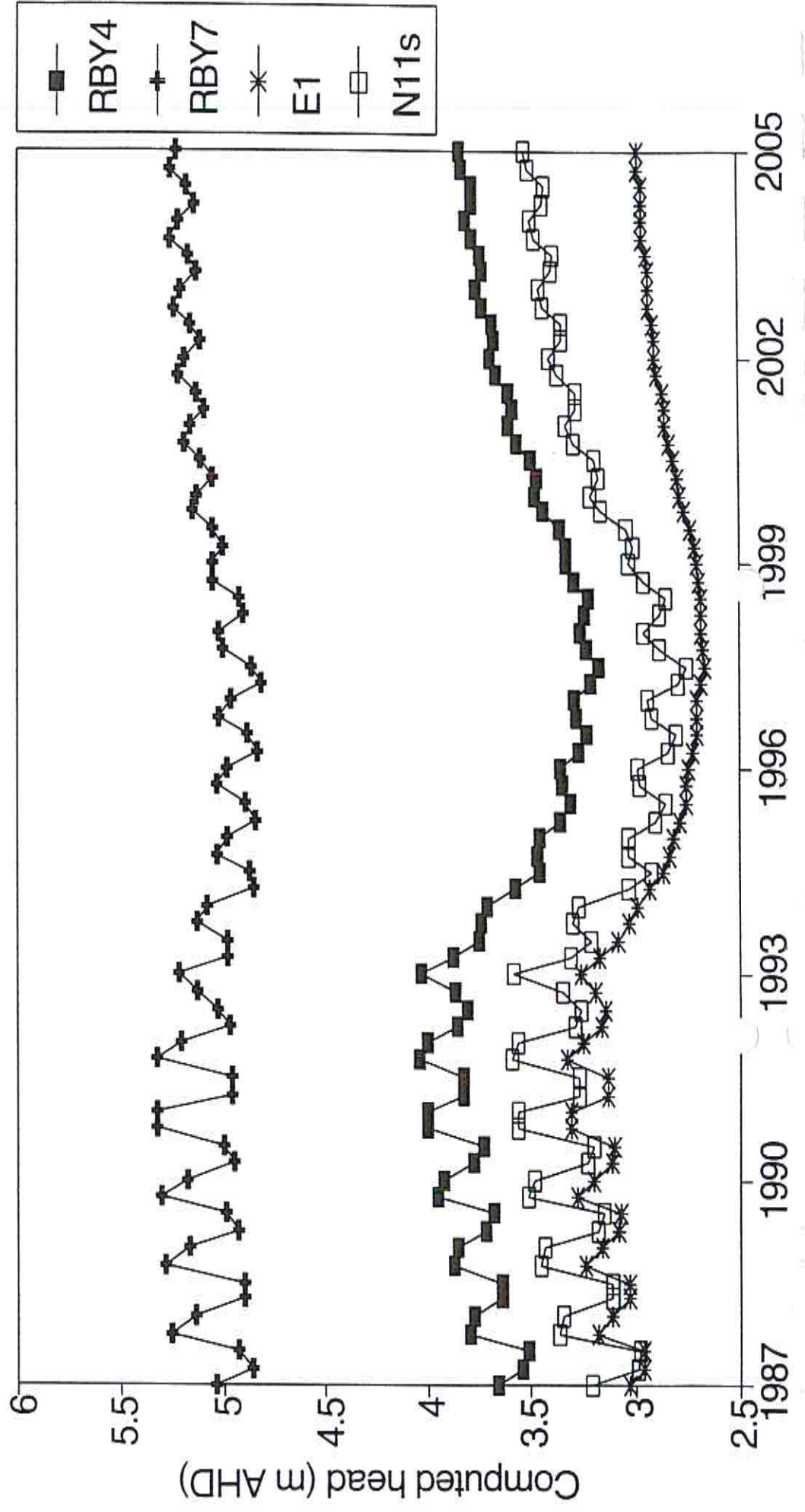


Figure 12: The impact of a reduction in recharge (with reduced discharge)



The final sensitivity analysis performed on the broad-mesh model involved a major modification to the type of upgradient (eastern) boundary condition used. Whilst a constant head boundary condition is reasonable over a short period of time, the impact of rising watertables in areas upgradient of the model would need to be considered for longer simulations. A general head boundary condition was used to simulate the case of heads rising in areas upgradient of the model at a rate of 0.05 m yr^{-1} , (ie. ranging from 6.0 to 6.85 m over the 18 year period).

In simulations the effect of the rising boundary condition was evident up to several kilometres from the boundary, although it was clear that the older discharge areas would short circuit the groundwater level rise for areas downgradient of them. The effect of this boundary condition is described in more detail in the fine-mesh model.

5.3 Broad-Mesh Model Conclusions

The broad-mesh model provided a useful insight into the dynamics of the groundwater system in the area. The dominance of vertical as compared to lateral flows became much more apparent through the use of the model. The model captured some of the seasonal trends in groundwater levels.

Boundary conditions were improved and an increasing head boundary condition was considered to be more appropriate than the constant head for the subsequent fine-mesh model in order to incorporate rises in groundwater levels due to the impact of areas upgradient of the model domain. In the western part and to the west of the model area, the occurrence of major discharge complexes have the capacity to absorb potential rises in hydraulic head. Measured heads from Barnett, (1992) indicate little long-term piezometric level variation in these areas. Therefore a constant head boundary condition was maintained on the western boundary.

The model was sufficiently detailed to obtain a reasonable match between computed and measured heads over the modelled period however the spatial discretisation of the grid was too crude for describing local flow systems, and also for giving confidence in the description of evaporative discharge and hence predicting longer term trends for land management options. The inability to predict the change in magnitude and extent of discharge due to changes in the proportion of land subject to salinisation (ie. subject to shallow watertables) required more detail at a finer scale in low-lying areas.

6. FINE-MESH MODEL

The fine-mesh model was developed to improve the manner in which localised processes (notably groundwater discharge) could be simulated, and provide a sounder basis for model calibration. This would then allow more confidence in applying the model to predict future groundwater levels for various land management options.

The fine-mesh model covered the same domain as the broad-mesh model, although it featured an improved level of discretisation. In its final form the model grid had dimensions of 59x47 (2773 cells). The dimensions of the cells were related to the topographic relief, and at their smallest were 25 metres within the study sites, wider around the discharge areas, and reached their maximum width near the model boundaries. Greatest resolution was given in areas which were currently discharging groundwater, or had the potential to become subject to evaporative discharge. The mesh for the entire study area is shown in Figure 13, whilst the more detailed mesh for sites 1 and 2 are shown in Figures 14 and 15 respectively. The details of all of the runs undertaken in the fine-mesh model are provided in Appendix 3.

6.1 Fine-mesh Steady State Calibration

The effort required for the steady state mode was minimised due to the previous effort with the broad-mesh model. Initial runs involved fine tuning of the grid. Following the generation of the final grid the steady state solution was generated as an initial condition for the transient modelling. The reliability of the solution was checked by comparison with the previous broad-mesh model and measured heads (Fig. 16). Model parameters were identical to those of the former model, with the treatment of recharge and discharge (including specified areas) almost identical for the two models. A very good agreement is evident for almost all piezometer sites, and slight discrepancies are attributed to differences of the grid.

6.2 Fine-mesh Transient Calibration

This section provides an overview of the approach undertaken to calibrate the fine-mesh model over the 6 year period and the results obtained. A wide range of parameters were tested in the transient mode in order to improve the performance of the model.

Piezometer hydrographs for 5 sites including site 1 (N11s), site 2 (E1), near the discharge lake (RBY10) and two other sites upgradient of the lakes (RBY4 and RBY7) are shown in Figure 17. They provide some necessary familiarisation of the piezometric response across areas of the model as they are referred to the following discussion.

During the transient simulations recharge rates were spatially uniform, but varied annually in accordance with rainfall (as shown in Table 2). This was an improvement over the broad-mesh model. For convenience in the text we refer to the mean recharge rates, rather than quote the range over the simulation period. Mean recharge was varied between 12 and 25 mm yr⁻¹.