HYDROSALINITY STUDIES IN THE COERNEY VALLEY

VOLUME 1: FINAL REPORT

J Herald

Institute for Water Research
Rhodes University

Report to the Water Research Commission on the project "Hydrosalinity studies in the Eastern Cape"

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The Steering Committee for this project requested the author to effect several corrections and amendments to the final report before recommending its acceptance by the Water Research Commission. For several reasons this was not done. In the interest of releasing potentially valuable research results to the discerning reader, the WRC has decided to release the draft report on special request.

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EXECUTIVE SUMMARY

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1. INTRODUCTION

The main objectives of this study were to gain a better conceptual understanding of the hydrosalinity processes of the lower Sundays River valley and to select and evaluate models or components of hydrosalinity models that are appropriate to irrigation management in the study area.

In part, the initial motivation for this research was provided by the Department of Water Affairs and Forestry who wished to investigate using the channel of the Sundays River as a conduit for the supply of Orange River water to the city of Port Elizabeth. However, at an early stage of the project it was decided to upgrade the main canal system of the Sundays River Irrigation Board and to use this infrastructure to carry water from Korhaans Drift to the new Scheepters Vlakte balancing dam. A pipe line from this dam now carries water to Port Elizabeth. This reduced the need for the study to focus specifically on the main lower Sundays River valley where data collection was difficult due to the extent of the area and the lack of adequate flow monitoring sites. At the meeting of the Steering Committee in March 1990 it was agreed that the study should concentrate on the lower Coerney valley and not the entire lower Sundays River valley. This decision did not substantially alter the initial aim of the project which was to gain a better conceptual understanding of the hydrosalinity processes within the lower Sundays River valley. However, it enabled the study to carry out a more detailed study of specific aspects of the hydrosalinity phenomena as related to irrigation, and irrigation return flows in particular.

The final report for this project is presented in two volumes. Volume 1 presents the methods and results of the main research programme in which a conceptual understanding of the hydrosalinity processes and an evaluation of several hydrosalinity models are presented. Volume 2 presents a summary of the data collected for the research discussed in Volume 1 and reports on the availability of this data.

2. RESEARCH AIMS

The specific objectives of the project were not adequately defined within the project proposal other than that the research should attempt to gain a better conceptual understanding of the hydrosalinity processes within the lower Sundays River valley and to select and evaluate models or components of models that are appropriate to the study area. However, to place this research into a more clear framework the following four objectives were identified :

i. To acquire data on the 3-D processes of moisture and solute movement at various spatial and temporal scales.

ii. To investigate and conceptually describe the hydro-chemical processes within the root and delivery zones.

iii. To test components of selected root zone and ground water models at various spatial and temporal scales.

iv. To determine the impact of irrigation on the streamflow and ground water within an area of irrigation.

The decision of the March 1990 Steering Committee, that this study should concentrate on the lower Coerney River and not investigate the entire lower Sundays River valley, did not substantially altered the initial aims of the project. In essence, this modification meant a change in the spatial scale at which the study was carried out.

3. RESEARCH PROCEDURE

To meet the objectives of this research project data were collected at the micro and catchment scale. Initially, micro-plots were established within two orchards to monitor the movement of soil moisture and solutes within the soil under an area of irrigated orchard. These data were used to examine the processes of soil moisture and solute movement within the root zone. Three management and one research level root zone hydrosalinity models were assessed in terms of their ability to simulate irrigation drainage. This assessment comprised sensitivity analyses and a comparison of the respective models predicted leaching flux values with those determined using the micro-plot data. At the catchment scale, streamflow and water quality monitoring sites were established and a series of boreholes drilled. Rainfall data were collected using a rain gauge network of 5 gauges and daily evaporation data were obtained from the Dept. of Agriculture's Addo Citrus Research Station. This facet of the study also investigated the use of stable isotopes for identifying the different water sources contributing to the hydrosalinity system of the lower Coerney valley. Information collected from this multi-scale data collection programme have provided a good conceptual understanding of the hydrosalinity processes that operate within the lower Sundays River valley and in particular within the Coerney area. Information on the weekly volume of water delivered to respective irrigation units within the Coerney area were provided by the Lower Sundays Irrigation Board. This information and that gained from a land use survey of the lower Coerney valley enabled a water balance for the area to be compiled.

4. THE LOWER SUNDAYS RIVER

The lower Sundays River valley lies some 40 km north of Port Elizabeth. For the purpose of this report the lower Sundays River valley is defined as the area extending from Korhaans Drift to Addo Bridge (figure 1.1). This area comprises an extensive area of intensive irrigation that comes under the control of the Lower Sundays River Irrigation Board.

The lower Sundays River is situated near the coastal belt and therefore receives rain throughout the year. Generally precipitation is higher in spring and autumn. There are large variations in the amount of rainfall due to the topographical characteristics of the area. In the mountainous areas rainfall exceeds 1100 mm per year while nearer the coast only 400 mm are received. The rain is chiefly cyclonic and thunderstorms are rare. The lower Sundays River valley has a mean annual rainfall of approximately 400 mm. The mean minimum temperatures in the month of July is 5° C while the mean maximum temperature in the month of January is 30° C. The mean annual A-pan evaporation for the area is approximately 1750mm.

The Witteburg Series of the Cape Super Group outcrop as predominant mountainous ridges to the north and east of the lower Sundays River valley. These intensively folded, hard rocks, comprise quartzite sandstones interbedded with thin layers of shale. The Uitenhage

Group comprising marine and fluvial laid cretaceous sedimentary rocks overlie the cape system and comprise the bed rocks of the Sundays River valley. More recent alluvial deposits over-lying the cretaceous mudstones form the current valley floor which is used for irrigation farming today. The irrigated soils in the lower Sundays River valley are mainly developed on alluvial and colluvial sediments. The dominant soil forms are Dundee, Oakleaf and Valsrivier. Irrigation farming was first introduced to the valley at around 1870 and in 1877 a Mr James

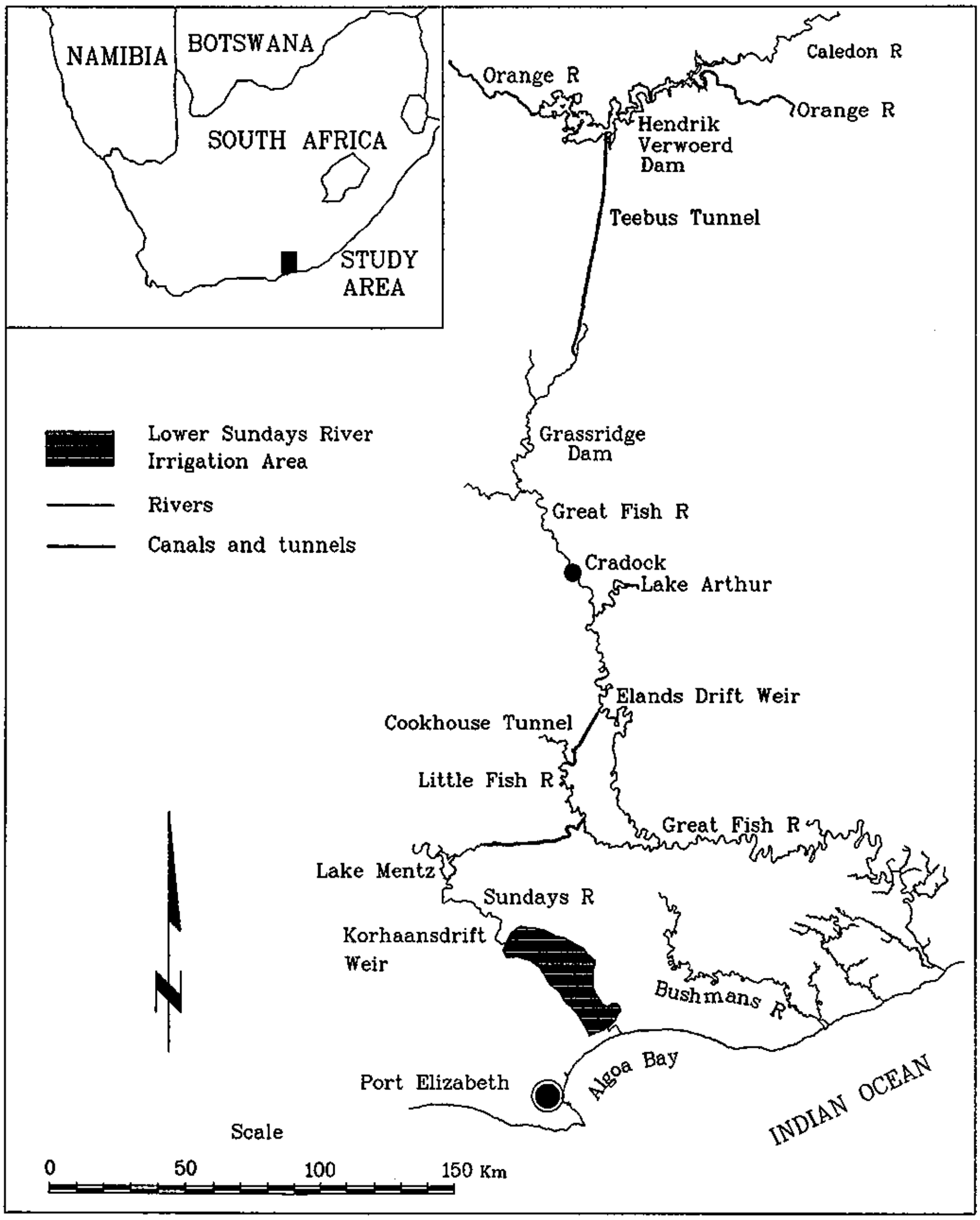

Figure 1.1 Location of the lower Sundays River irrigation area and its link to the Orange River scheme.

Kirkwood attempted to form an irrigation co-op. However, it was not until the construction of the Lake Mentz dam in 1922 that large scale irrigation began to take place within the valley. The wall of Lake Mentz dam was raised by 1.5 m in 1935 and again in 1952 by 5.8 m. However during periods of drought the valley was still subject to periodic water shortages. Areas that had been originally scheduled for irrigation, but had not been developed were therefore de-scheduled in order to create a more assured yield for the rest of the valley. In spite of these measures the valley was still subject to periods of water shortage until 1978 when Lake Mentz was linked to the Orange River Scheme (figure 1.1). The valley is now assured of an adequate water supply, although at times the salinity level has risen above normally acceptable levels.

The Sundays River valley produces mainly navel and valencia oranges, lemons and other loose skinned citrus fruit, as well as lucerne and potatoes. Other types of farming such as sheep, cattle and game farming take place to a lesser extent. In 1985, 9028 ha of farmland were scheduled for irrigation, while another 3198 ha were identified for future expansion. The regional importance of this irrigation development is clearly shown by its 1983 economic returns of R19.9 million in citrus exports and R2 million to locally markets.

5. CONCLUSIONS

Initially the study examined variations of soil moisture, and soil water and solute flux within the root zone. Data were collected at a number of micro-plots, but only that for one plot, DDM03, located in orchard M of Daisy Dell farm was used for this investigation. The most important conclusion drawn from this aspect of the study was the importance of macro-pore flow. Soil moisture flux was initially determined using a Darcian approach which is driven by the difference in matrix potential between two points in the soil profile. This method, which can only account for micro-pore flow was found to grossly underestimate the total soil water flux. A water balance approach was found to provide more realistic results, but was discarded due to its gross over estimation of negative fluxes when the evaporative term exceeded available soil moisture. The study clearly demonstrates how the water balance approach can not be implemented using a fixed crop factor for determining the evaporative term. To more successfully implement this approach one would require the knowledge needed to more accurately define the transpiration response of citrus trees to variations in the factors controlling the evaporative processes. Therefore, due to problems related to both the Darcian and the water balance methods of determining irrigation drainage a third approach based on a chloride mass balance was developed. This approach provided the most realistic estimate of soil moisture flux. The soil moisture fluxes determined by this approach were very similar to those determined by the water balance approach, but without the large negative fluxes between irrigation events.

A comparison of the soil moisture fluxes as determined using the Darcian approach, which can only account for micro-pore flow, and those determined using the mass balance approach, clearly indicated the importance of macro-pore flow as a component of irrigation drainage. The chloride mass balance approach was subsequently used to determine the solute flux within the root zone of the study orchard.

The third objective of the study was to evaluate a number of root zone models. Initially it was decided by the steering committee for this project, that this aspect of the study should rely on the finding of a another Water Research Commission funded study which set out to evaluate a number of hydrosalinity models for the specific purpose of estimating soil water and solute movement within the root zone (Moolman, 1991). However, as the progress of that study was delayed, this project undertook a more independent approach. Also, as this study's investigation of soil moisture fluxes clearly indicated the importance of macro-pore flow, it became apparent that the models being examined by Moolman (1991) were inappropriate to the study area. In fact, from a study of the literature, it appears that most hydrosalinity models are based on the Richards equation which can only account for micro-pore flow. It was therefore decided to examine a number of hydrosalinity models, knowing full well that they would prove inappropriate to modelling irrigation drainage within the study area. The main objective being to clearly demonstrate the need for research into the development of models suitable for macro-pore flow dominated soil moisture drainage. Three management and one research type model were selected to cover the full range of model complexities. As expected the output from all four models, when compared with the soil moisture fluxes determined by the mass balance approach, were very disappointing. These results are of great significance as they clearly demonstrate the need for hydrosalinity models that are applicable in soils where macro-pore flow is the dominant form of soil moisture drainage. Within South Africa, which is faced with ever increasing water resource limitations and where over 50 percent of this resource is used for irrigation agriculture, there is an urgent need for research into, and improved management of, irrigation farming. The development of applicable hydrosalinity research and irrigation management models is an area of research that should be given a far greater priority than is currently the situation.

The forth objective of the study was to determine the impact of irrigation on both the ground water and streamflow within the study area. A number of boreholes were drilled by the Department of Water Affairs and Forestry. The level and salinity of the ground water aquifer within the alluvial deposit of the lower Coerney valley were monitored at these sites. The vertical salinity profile of the aquifer was also recorded at each borehole. These data clearly indicated a marked rise in the level of the water table in response to irrigation farming. The area of influence extended some distance up valley from irrigation development. It was also learnt that the salinity of the irrigation drainage is significantly less than that of the natural ground water within the valley. An isolated pocket of highly saline water was very clearly indicated by constructing an isoline map of electrical conductivity values recorded within the lower Coerney valley. The salinity of this water is very similar to that of the aquifer further up valley from the area influenced by irrigation. An examination of the vertical salinity profile at this point indicated that the highly saline water extended from the upper to the lower surface of the aquifer. In close proximity to this area of high salinity a shallow borehole drilled through the alluvium and into the underlying cretaceous mudstones struck artesian water. This suggests that artesian water is rising up through the underlying marine laid mudstones from the even deeper Table Mountain Sandstones. After moving up through the marine laid mudstones the relatively less saline water of the Table Mountain Sandstone has an electrical conductivity in excess of 2000 $mS.m^{-1}$. The less saline irrigation drainage water initially sits above the more dense, very saline, natural ground water. The salinity of this aquifer is therefore highly variable with both lateral and vertical gradients. This understanding clearly indicated the dangers of monitoring the surface water of an aquifer in the hope of determining its water quality characteristics.

The study also examined the discharge and salinity regimes of the lower Coerney River itself. Discharge records clearly indicate a steady rise in the base flow of the river as irrigation farming expands within the valley. Simultaneously the solute concentration of the river has decreased as the natural ground water component of the streamflow is diluted with an increasing proportion of less saline irrigation return flow. Stable isotopes were successfully used to determine the natural ground water and irrigation return flow components of both streamflow and local ground water at various locations within the study area. This technique was found to provide more useful information than conventional water quality analyses for gaining an understanding of the hydrological and hydrochemical systems of the study area.

6. RECOMMENDATIONS FOR FUTURE RESEARCH

The current study was carried out at a number of scales from micro-plot studies to an examination of the impact of irrigation on ground water and streamflow at the catchment scale. Analyses of the data collected at these scales and conclusions drawn from an assessment of several hydrosalinity models has highlighted a number of areas for future research:

i. To establish a methodology for determining the magnitude and spatial variability of macro-pore flow.

ii To carry out detailed studies of the hydrosalinity processes within the delivery zone at the plot scale rather than at the catchment scale.

iii. To gain a better understanding of the spatial variations of hydrosalinity processes at the field scale.

iv. To develop a hydrosalinity research model that considers macro-pore flow and accounts for spatial variability at the field scale.

v. To develop an irrigation systems model that permits the movement of ground water between adjacent ground water cells and allows for vertical salinity gradients within the local ground water system.

Clearly this study highlighted the importance of macro-pore flow as a dominant process in irrigation drainage. However, there is currently no sound methodology by which this component of soil water movement can be readily estimated and accounted for in hydrosalinity models. Until this aspect of hydrosalinity modelling is further understood it is difficult to foresee any significant progress in the field of irrigation management which is so desperately needed in this country of limited water resources.

Another very problematic area of salinity research lies with the lack of understanding of the processes operating within the delivery zone. This zone which connects the root zone to the ground water and surface water systems is very difficult to monitor because of its depth and also because it is often comprised of cobble and bolders which are difficult to penetrate with drilling equipment which is appropriate to research studies. Yet, being an important link for the transfer of irrigation drainage to the ground water and streamflow systems, it is important to understand how solutes may either be accumulated and leached from this zone if the impacts of an irrigation development are to be properly assessed.

For the proper extrapolation of information gained from plot studies of the hydrosalinity processes, it is important that one gains an understanding of the spatial variability of these processes. If one can statistically define these variations it may be possible to develop more stochastic type models and reduce the data requirements of spatially distributed, deterministic type hydrosalinity models.

Currently, due to the increasing pressure on South Africa's water resources there is a great need for improved irrigation efficiency. If one considers the large volume of water currently used for irrigation, it is clear that even very small improvements in irrigation efficiency may lead to relatively large savings in water. There is also an increasing need to reduce the pollution of surface water resources by irrigation drainage. For example, the salinity of water transferred via the Orange/Fish River scheme increases in salinity from approximately 35 $mS.m^{-1}$ at Katkop (Q1M01) to over 75 $mS.m^{-1}$ at Elands Drift (Q5R0101), a distance of less than 50 km. This increase in salinity is largely attributed to saline irrigation return flow generated along this reach of the Great Fish River. The cost of this irrigation practice in terms of loss of production in the lower Sundays River irrigation area has not been estimated, but it is most probably considerable. Appropriate hydrosalinity models would provide an ideal methodology for determining more efficient irrigation strategies with the potential of saving water and reducing pollution. However, before selecting a model for this work one would need to ensure that it was appropriate to the conditions of the study area. In particular, the importance of macro-pore flow must be determined and if necessary an appropriate model developed. Currently it would appear that such a model is not available and that research into this aspect of hydrosalinity modelling is essential.

To evaluate the impact of an irrigation development on either the ground or surface water resources of an area, an appropriate irrigation management systems model is required. Such a model has been developed for the Breede River. However, as the structure of this model was developed specifically for the Breede River valley it does not facilitate the movement of ground water between adjacent cells. It also does not facilitate variations of salinity in the vertical profile of the aquifer and recharge of very saline water from an underlying artesian aquifer. It would seem that to meet these more flexible modelling requirements a finite difference type model would be required. Such a model would have wider application than the currently available model developed for the Breede River irrigation area.

Further to this list of specific hydrosalinity research requirements, it is recommended that the monitoring and analyses of data collected for the lower Coerney valley, initiated in this study, should be continued. There is now an historical record of the impact of irrigation on both the ground water and streamflow systems within the valley. There is also a good conceptual understanding of the hydrological system and methods for identifying components of this system have been established. Currently irrigation farming within the lower Coerney valley is under rapid expansion following the construction of a new higher level canal. It would therefore make a great deal of sense for at least some of the recommended areas of research, as listed above, to be carried out in this study area. It is only through continuity of clear research objectives that progress in such a problematic, yet important field as hydrosalinity research that progress can be made.

7. REFERENCES

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LIST OF ABBREVIATIONS

For ease of presentation the water quantity and quality parameters summarise in this volume are often abbreviated with the following abbreviations.

Abbreviation	Parameter	Units
WL	Water level	m
Q	Discharge	$m^3.s$
MPot	Matric potential	kPa
pH	Concentration of H^+ ions	
EC	Electrical conductivity	$mS.m^{-1}$
TDS	Total dissolved solids	$mg.l^{-1}$
Talk	Total alkalinity ($CaCO_3$)	$mg.l^{-1}$
Cl	Chloride	$mg.l^{-1}$
Ca	Calcium	$mg.l^{-1}$
K	Potasium	$mg.l^{-1}$
Mg	Magnesium	$mg.l^{-1}$
Na	Sodium	$mg.l^{-1}$

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I am also greatly indebted to the farming community of the lower Sundays River valley, and in particular Mr John Hanah of the farm Daisy Dell and his manager Mr Boetie Classen. Without their assistance the micro-plot aspect of this study would not have been possible.

The contribution of Mr Haremse du Plessis of the Lower Sundays Irrigation Board who made irrigation scheduling data available and provided a contact with the farming community is also greatly appreciated.

I would also like to express my thanks to the Department of Water Affairs and Forestry for the construction of the flow monitoring sites and drilling a number of boreholes within the lower Coerney valley. The data collected at these sites was crucial to the success of this project.

Finally I would like to thank the following people who comprised the Steering Committee for their contribution to this project. Mr Meiring du Plessis deserves a special thanks for his guidance as responsible Water Research Commission adviser.

Mr H.J. du Plessis	Water Research Commission
Mr J.H. du Plessis	Sundays Irrigation Board
Prof J.H. Moolman	Stellingbosch University
Mr A.G. Reynders	Water Research Commission
Dr A.J. vd Merwe	Department of Agriculture
Mr M. van Veelan	Dept. of Water Affairs and Forestry
Mr F.P. Marais	Water Research Commission

1. INTRODUCTION

This volume is one of three reports that deal with the results of the Water Research Commission funded project "Hydrosalinity studies in the eastern Cape", undertaken by the Institute for Water Research at Rhodes University. The main objectives of this study were to gain a better conceptual understanding of the hydrosalinity processes of the lower Sundays River valley and to select and evaluate models or components of hydrosalinity models that are appropriate to irrigation management in the study area.

In part, the initial motivation for this research was provided by the Department of Water Affairs and Forestry who wished to investigate using the channel of the Sundays River as a conduit for the supply of Orange River water to the city of Port Elizabeth. However, at an early stage of the project it was decided to upgrade the main canal system of the Sundays River Irrigation Board and to use this infrastructure to carry water from Korhaans Drift to the new Scheepters Vlakte balancing dam. A pipe line from this dam now carries water to Port Elizabeth. This reduced the need for the study to focus specifically on the main lower Sundays River valley where data collection was difficult due to the extent of the area and the lack of adequate flow monitoring sites. At the meeting of the Steering Committee in March 1990 it was agreed that the study should concentrate on the lower Coerney River valley and not the entire lower Sundays River valley. This decision did not substantially alter the initial aim of the project, which was to gain a better conceptual understanding of the hydrosalinity processes within the lower Sundays River valley. However, this enabled the project to carry out a more detailed study of specific aspects of the hydrosalinity phenomena as related to irrigation, and irrigation return flows in particular. The study was therefore able to focus in more detail on the processes at the micro-scale, examine a number of hydrosalinity models and to evaluate the use of geophysical techniques for gaining a greater knowledge of the alluvial deposits and their hydraulic properties.

1.1 RESEARCH AIMS

The aims and objectives of the original research proposal were:

> *"The proposed research aims to contribute towards a lack of available predictive tools for the management of South Africa's irrigation schemes. This may be achieved by conducting detailed geohydrological investigations of delivery zones and conceptualising the active hydro-chemical processes. Appropriate models or components of ground water movement and hydro-chemical processes could then be selected and coupled with existing root zone models and flow routing methods. To enable practical application of the results from this study the research will seek to develop simple, rather than complex modelling strategies. Ideally, the research should be conducted in an irrigation area where,*
>
> *(i) data collection is already active,*
>
> *(ii) water resource planners require information on irrigation return flow quantities and qualities,*

(iii) the effects of expanding irrigation on return flow quantities and qualities can be assessed.

It is proposed that the research should be conducted at various spatial and temporal scales in the Lower Sundays River Valley. While the objectives are viewed as a whole it is nevertheless convenient to split the objectives into the following four related components.

i. To investigate and conceptually describe the hydro-chemical processes within the delivery zone of selected areas at various spatial and temporal scales in the Lower Sundays River Irrigation Area.

ii. To acquire data on the 3-dimensional processes of moisture and solute movement within selected areas at various spatial and temporal scales in the Lower Sundays River Irrigation Area.

iii. To test components of individual root zone and ground water models at various spatial and temporal scales and to combine selected models or components to simulate the quantity and quality of irrigation return flows.

iv. To simplify the model as much as possible to facilitate its use as a management tool at the macro-scale. Simplification will aim at minimising the input requirements and improving the computational efficiencies".

If one considers these "four related components" as the specific objectives of this study then, given the manpower and resources available, the aims of this project were somewhat ambitious. The proposed research framework was correct in that it initially sought to investigate and conceptualize the hydrosalinity processes affecting the quantity and quality of water within the root and delivery zones. This was to be done at both the plot and catchment scale. The proposed methodology included the establishment of micro-scale plots to examine soil water and solute transport in the root zone. However, to acquire and install the necessary equipment and then to collect and analyse these data was a more time consuming task than originally envisaged. Having determined the factors controlling the processes of irrigation return flow, it was then proposed to compile a hydrosalinity model using components from existing models. This hybrid model was to be adapted to provide a more simple model that may be used as a management tool for predicting the impact of irrigation on return flow quantities and qualities at the catchment scale. The objectives as specified in the proposal were unobtainable within the permissible time frame.

As well as the micro-plot studies the proposed methodology also included monitoring the inputs and outputs from 1 to 10 km^2 cells within an irrigated alluvial deposit. It also suggested that transects of boreholes should be established, connecting the micro-plot sites to the river, thereby enabling the linkage between irrigation applications at the micro-plot and the emergence of irrigation return flow in the river to be monitored. This methodology assumes that one can delineate and monitor cells within an alluvial aquifer and that the ground water flow is orthogonal to the stream channel. In fact, ground water flow within an alluvial deposit is generally in a down valley direction, but with many local variations due to the heterogeneity of the alluvial deposits hydraulic characteristics. The methodology was therefore based on conceptual misconceptions and could not be successfully implemented within an alluvial aquifer like that of the lower Sundays River valley.

Clearly the specific objectives and the proposed methodology of the project were not well defined within the project proposal other than that the research should attempt to gain a better conceptual understanding of the hydrosalinity processes within the lower Sundays River valley and to select and evaluate models or components of models that are appropriate to the study area. Therefore, to place this research into a more clear framework the following more realistic objectives were identified :

i. To acquire data on the 3-D processes of moisture and solute movement at various spatial and temporal scales.

ii. To investigate and conceptually describe the hydro-chemical processes within the root and delivery zones.

iii. To test components of selected root zone and ground water models at various spatial and temporal scales.

iv. To determine the impact of irrigation on the streamflow and ground water within an area of irrigation

The decision of the March 1990 Steering Committee, that this study should concentrate on the lower Coerney River and not investigate the entire lower Sundays River valley, did not substantially alter the initial aims of the project. In essence, this modification meant a change in the spatial scale at which the study was carried out.

Initially it was proposed that the study would be carried out at three scales; the micro, meso and macro-scales. The micro-scale referred to plot studies, the meso-scale to the Coerney area and the macro-scale to the entire lower Sundays River valley. It was then decided that the different scales should refer to: the micro-scale as the processes at a point in space, the meso-scale as the field scale and the macro-scale as the lower Coerney River irrigation area. These definitions of scale are possibly more appropriate than those of the initial research proposal in terms of the scales at which the results may be applied for management purposes.

1.2 RESEARCH PROCEDURE AND REPORT LAYOUT

To meet the objectives of this research project, data were collected at the micro and catchment scale. Initially micro plots were established within two orchards to monitor the movement of soil moisture and solutes within the soil under an area of irrigation. At the catchment scale streamflow and water quality monitoring sites were established and a series of boreholes drilled. Rainfall data were collected using a rain gauge network of 5 gauges and daily evaporation data were obtained from the Dept. of Agriculture's Addo Citrus Research Station. Information collected from this multi-scale data collection programme has provided a good conceptual understanding of the hydrosalinity processes that operate within the lower Sundays River valley and in particular within the Coerney area. Information on the weekly volume of water delivered to respective irrigation unit within the Coerney area were provided by the Lower Sundays Irrigation Board. This information and that gained from a land use survey of the lower Coerney valley enabled a water balance for the area to be compiled.

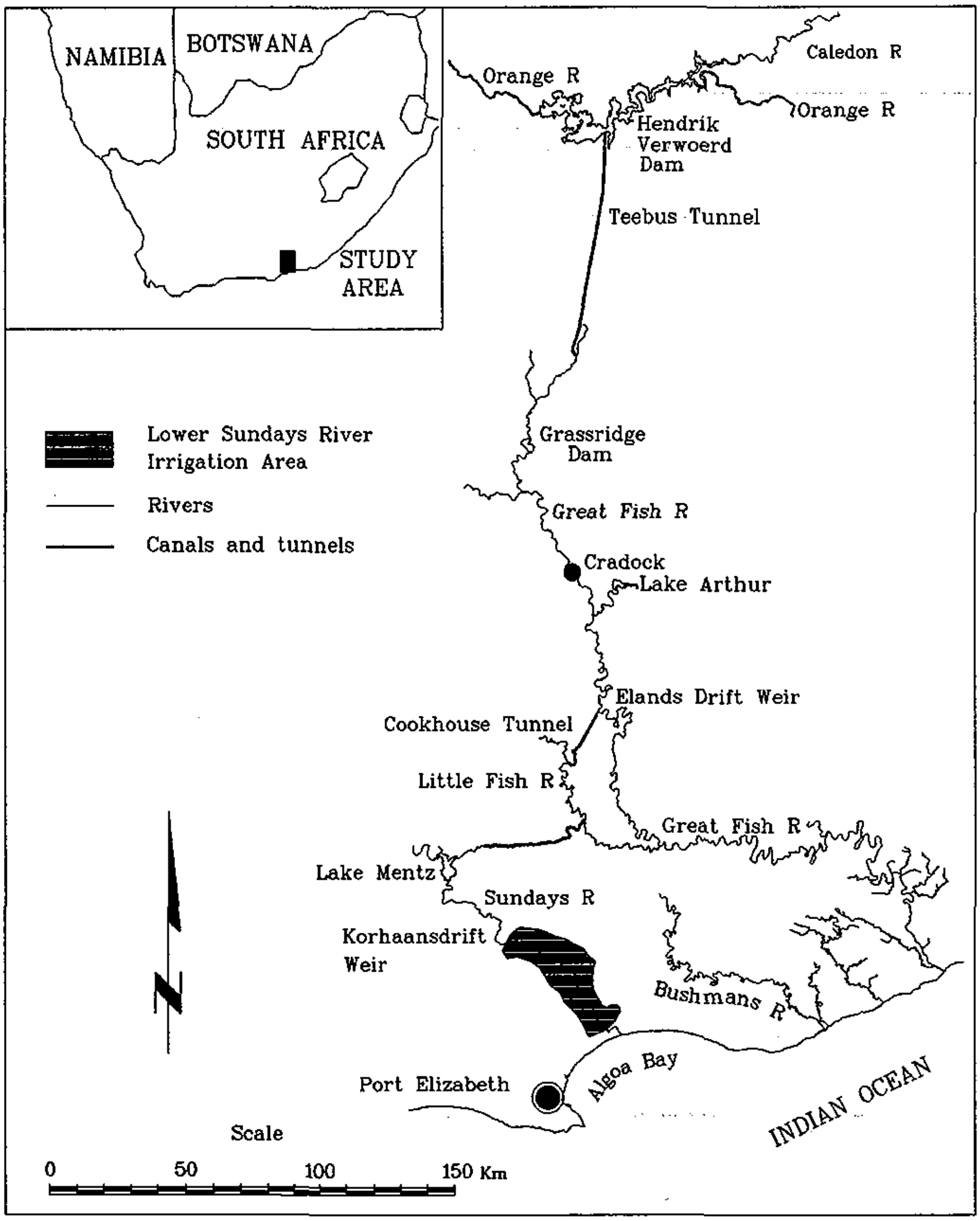


Figure 1.1 Location of the lower Sundays River irrigation area and its link to the Orange River scheme.

The final report for this project is presented in three volumes. Volume 1 presents the methods and results of the main research programme in which a conceptual understanding of the hydrosalinity processes and an evaluation of several hydrosalinity models were carried out. Volume 2 presents a summary of the data collected for the research discussed in Volume 1 and reports on the availability of this data. An MSc thesis from Rhodes university presents the finding of a two year sub-project which evaluated the use of several geophysical techniques for delineating buried stream channels within alluvial deposits.

The first three chapters of this volume introduce the research topic and the study area as well as providing a theoretical background to the processes and phenomena most relevant to this field of research. The fourth and fifth chapters examine the hydrosalinity process of soil water and solute movement at the plot scale and the modelling of these processes. Chapters 6, and 7 examine hydrosalinity process and the impacts of irrigation on the ground water and streamflow components of the hydrological cycle. Chapter 8 describes an attempt to implement a systems model to the Coerney River valley which was not completed. Finally, Chapter 9 summarises the findings of this project and makes recommendations for future research.

1.3 THE LOWER SUNDAYS RIVER VALLEY

The lower Sundays River valley lies some 40 km north of Port Elizabeth and for the purpose of this report is defined as the area extending from Korhaans Drift to Addo Bridge (figure 1.1). This area comprises an area of intensive irrigation that comes under the control of the Lower Sundays River Irrigation Board.

The lower Sundays River is situated close to the Eastern Cape coastal belt and therefore receives rain throughout the year. Generally precipitation is higher in spring and autumn. There are large variations in the amount of rainfall due to the topographical characteristics of the area. In the mountainous areas rainfall exceeds 1 100 mm per year while nearer the coast only 400 mm are received. The rain is chiefly cyclonic and thunderstorms are rare. The lower Sundays River valley has a mean annual rainfall of approximately 400 mm. The mean minimum temperature July is 5° C while the mean maximum temperature in January is 30° C. The mean annual A-pan evaporation for the area is approximately 1 750mm (Tylcoat 1985).

The Witteburg Series of the Cape Super Group outcrop as predominant mountainous ridges to the north and east of the lower Sundays River valley. These intensively folded, hard rocks, comprise quartzite sandstones interbedded with thin layers of shale. The Uitenhage Group comprising marine and fluvial laid cretaceous sedimentary rocks overlie the cape system and comprise the bed rocks of the Sundays River valley. More recent alluvial deposits over-lying the cretaceous mudstones form the current valley floor which is used for irrigation farming today. A more detailed description of the regional geology and alluvial deposits is presented in the description of the Coerney valley (Chapter 2). The irrigated soils in the lower Sundays River valley are mainly developed on alluvial and colluvial sediments. The dominant soil forms are Dundee, Oakleaf and Valsrivier. A more detailed description of soils within the lower Coerney valley is presented in Chapter 2.

Irrigation farming was first introduced to the valley in 1870 and in 1877 a Mr James Kirkwood attempted to form an irrigation cooperative. However, it was not until the construction of the Lake Mentz dam in 1922 that large scale irrigation began to take place within the valley. The wall of Lake Mentz dam was raised by 1.5 m in 1935 and again in 1952 by 5.8 m. However during periods of drought the valley was still subject to periodic

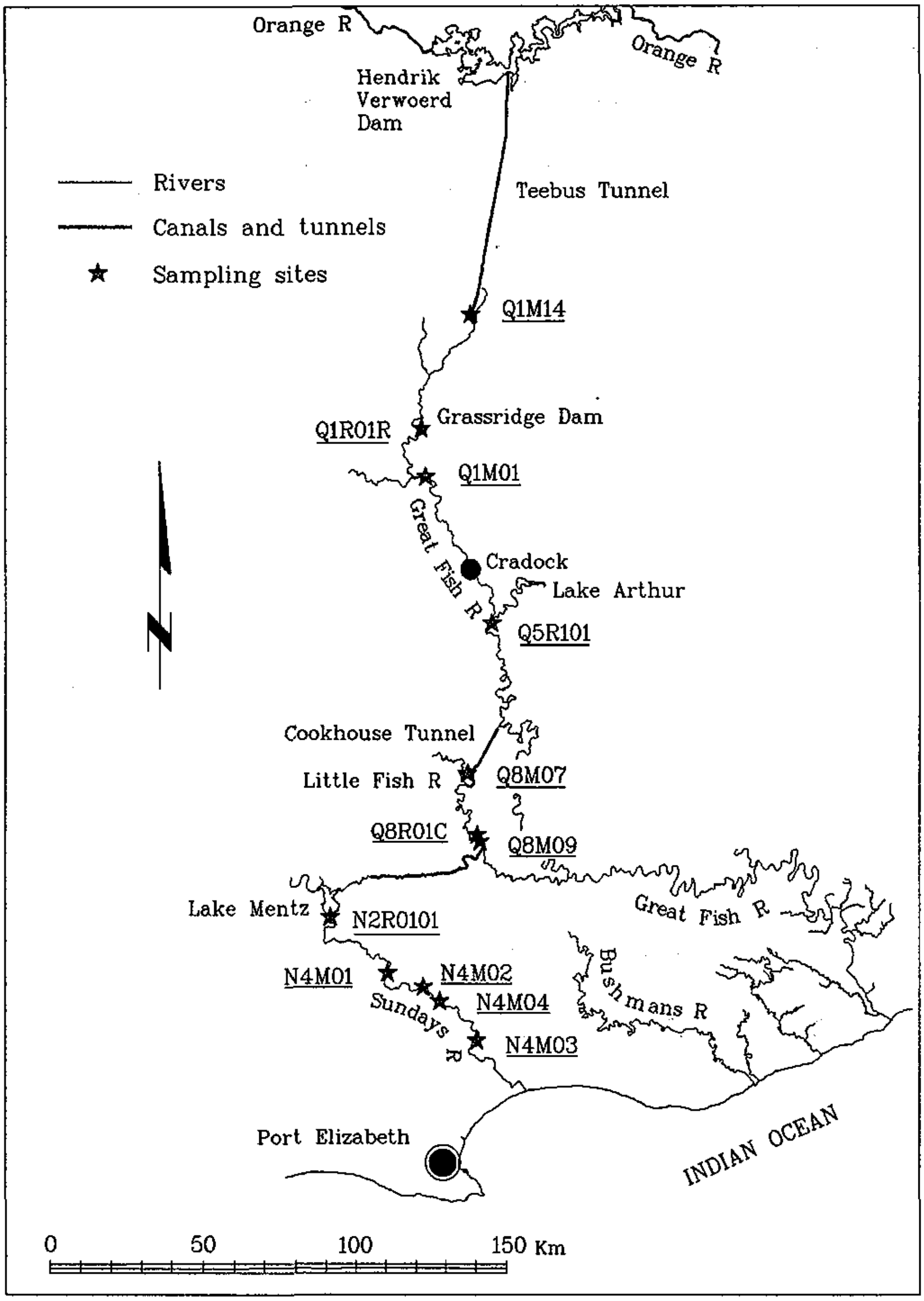


Figure 1.2 Water quality monitoring sites along the Orange/Fish water scheme serviced by the Department of Water Affairs and Forestry.

water shortages. Areas that had been originally scheduled for irrigation, but had not been developed were therefore descheduled in order to create a more assured yield for the rest of the valley. In spite of these measures, the valley was still subject to periods of water shortage until 1978 when Lake Mentz was linked to the Orange River Scheme (Figure 1.1). The valley is now assured of an adequate water supply, although at times the salinity level has risen above normally acceptable levels.

The Sundays River valley produces mainly navel and valencia oranges, lemons and other loose skinned citrus fruit, as well as lucerne and potatoes. Other types of farming such as sheep, cattle and game farming take place to a lesser extent. In 1985, 9 028 ha of farmland were scheduled for irrigation, while another 3 198 ha were identified for future expansion. The regional importance of this irrigation development is clearly shown by its 1983 economic returns of R19.9 million in citrus exports and R2 million to local markets.

1.4 SALINITY OF SUNDAYS RIVER IRRIGATION WATER

In 1978 Lake Mentz was linked to the Orange River system which feeds water from the Hendrik Verwoerd Dam, on the Orange River, through the Teebus Tunnel to the Great Fish River. After flowing down the channel of the Great Fish River for approximately 200 km the water is diverted through the Cookhouse Tunnel to the Little Fish River. Then from the Wellington Grove pump station, on the Little Fish, water is transferred via a canal to the Skoenmakers River which flows into Lake Mentz. Unfortunately, as the water passes through this system, especially along the river channel of the Great Fish River from Katkop (Q1M01 - DWAF old numbering system) to Elands Drift (Q5R0101), the electrical conductivity of this water increases from a median value of approximately 33 to 94 $mS.m^{-1}$ (figures 1.2 and 1.3). The concentration of the chloride ion increases from 27 to 124 $mg.l^{-1}$.

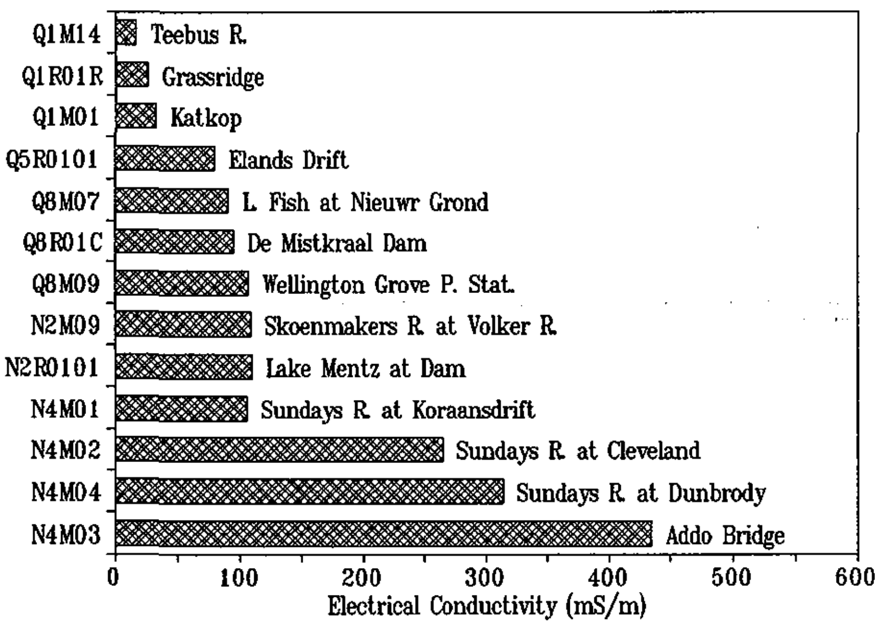


Figure 1.3 Median values of electrical conductivity of water at Department of Water Affairs and Forestry monitoring sites along the Orange/Fish water scheme.

This increase in salinity is attributed to the saline irrigation return flow generated along this reach of the river. Water is pumped from the river to irrigate lands along its banks. A proportion of this water leaches through the soil and returns to the river with an increased concentration of soluble salts. This leads to a deterioration of the water moving through the Orange/Fish Water Scheme, such that by the time it reaches the irrigation areas of the lower

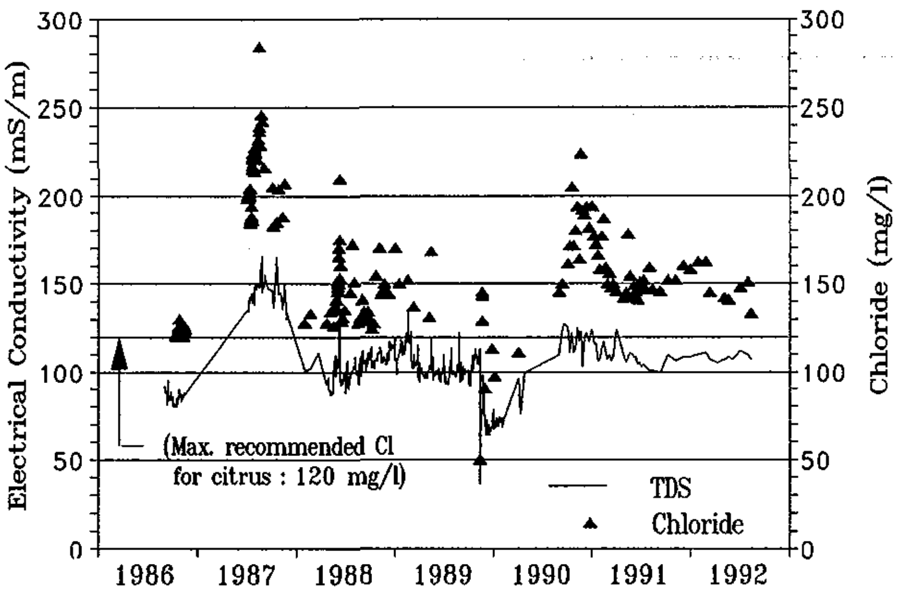


Figure 1.4 Variations of electrical conductivity and chloride concentrations in Korhaansdrift for the period 1986 to 1992.

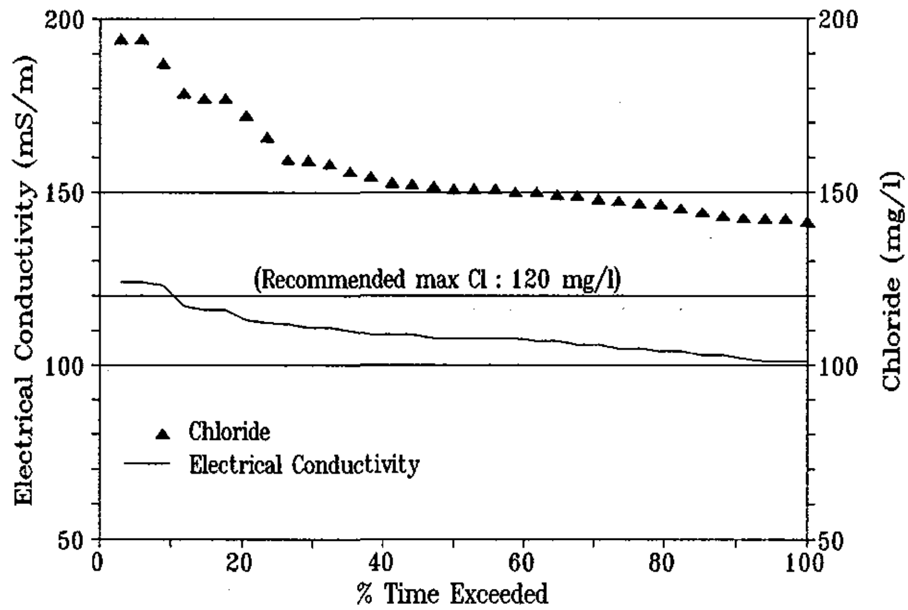


Figure 1.5 Percentage time for which levels of electrical conductivity and chloride concentrations were exceeded during 1990 for water in Korhaansdrift.

Fish and Sundays the salinity has increased to such a level that it can no longer be considered ideal for irrigation. Within the current literature the salt tolerance of citrus, the predominant irrigation crop of the lower Sundays River irrigation area, is not clearly defined in terms of either the permissible electrical conductivity or concentration of the chloride ion. It is not sufficient to adopt information from other areas as the permissible salinity of irrigation water. This can only be determined from a careful study of a particular crops tolerance, the soil permeability and the method of irrigation application. Although the generally accepted permissible maximum concentration of chloride in irrigation water for successful citrus production is 340 mg l^{-1}, du Plessis (1975) recommends that for the lower Sundays River area the level of chlorides should not exceed 120 mg l^{-1}. Figures 1.4 and 1.5 clearly show that irrigation water delivered from the Orange Fish River Scheme often exceeds du Plessis' recommendation. Figure 1.4 shows variations of both EC and chloride in irrigation water passing through Korhaansdrift for the period 1986 to 1992. Clearly there are significant temporal variations in the salinity levels of irrigation water with the chloride level rising significantly above that recommended for use in the lower Sundays River irrigation area. The temporary decrease in salinity at the end of 1989 was due to a major flood event that flushed Korhaansdrift with natural runoff and reduced the salinity to approximately 34 $mS.m^{-1}$. Figure 1.5 illustrates the percentage time for which levels of salinity and chloride were exceeded during the 1990/91 irrigation season. Not only was the recommended maximum level of chloride exceeded for the entire period, but for more than 50 % of the season the chloride level exceeded 150 mg l^{-1} and at times rose to above 190 mg l^{-1}. To assess the implications of using relatively saline water for irrigation in terms of crop production and irrigation management are beyond the scope of this report. However, from economic considerations alone this situation should be investigated further, especially as such a large proportion of the salt load of the Orange River water originates as irrigation return flow from the short stretch of channel between Katkop and Elands Drift.

1.5 SIGNIFICANCE OF RESEARCH

The Sundays River irrigation area has been subject to problems of salinization and water logging of the soil under irrigation from as early as 1930. As far back as 1927 farm mis-management was blamed for impaired drainage and declining soil quality (Shone, 1976). In 1981 the problems of chlorosis, defoliation, die-back, unseasonal fruit set, root disease, varying yields and quality deterioration among others led to an attempt to establish the causal factors. Physical characteristics of the soil were identified as responsible for poor drainage. Management practises such as orchard traffic and extensive levelling increased soil compaction, and heavy irrigations then exacerbated the problems. These conditions need light and frequent irrigation applications, and hence a finer resolution of control over the amount and timing of water application. The greater level of control required by farmers is intended to steer a course between water logging the soil and excessive salt accumulation. There must however be a basis for management decisions. It is towards satisfying this need that the current research makes a worthwhile contribution. Until the factors controlling water and solute movement within the root zone of the study area are properly identified, correct management procedures cannot be determined.

Another factor adding to the significance of this study is the scheduled expansion of the irrigation area within the lower Sundays River valley from the current extent of approximately 9000 ha by over 30% to 12200 ha (Tylcoat, 1985). By monitoring the impact of this land-use change on both the ground and surface water systems an invaluable record will be compiled for the future development and testing of hydrosalinity and irrigation management systems models.

2. THE LOWER COERNEY RIVER VALLEY

The lower Coerney River valley is situated approximately 40 km north of Port Elizabeth between latitudes 33° 22' and 33° 31' south and longitude 25° 35' and 25° 45' east. This area comprises the catchment area of the Coerney River from its confluence with the Sundays River to where the road running between Coerney and Zuurberg crosses the river (figure 2.1). The area was selected as a definable sub-area of the lower Sundays River valley, which is important for its current citrus production and includes a significant area planned for future irrigation development. This planned development should result in the area currently under irrigation within the lower Coerney River valley increasing from approximately 792 ha to over 1 700 ha during the next ten years. This increase comprises nearly a third of the planned expansion for the entire lower Sundays River area following completion of the new higher level canal. Therefore, by establishing a monitoring programme within this valley an opportunity is created whereby it should be possible to determine the long term impact of irrigation within the lower Sundays River valley.

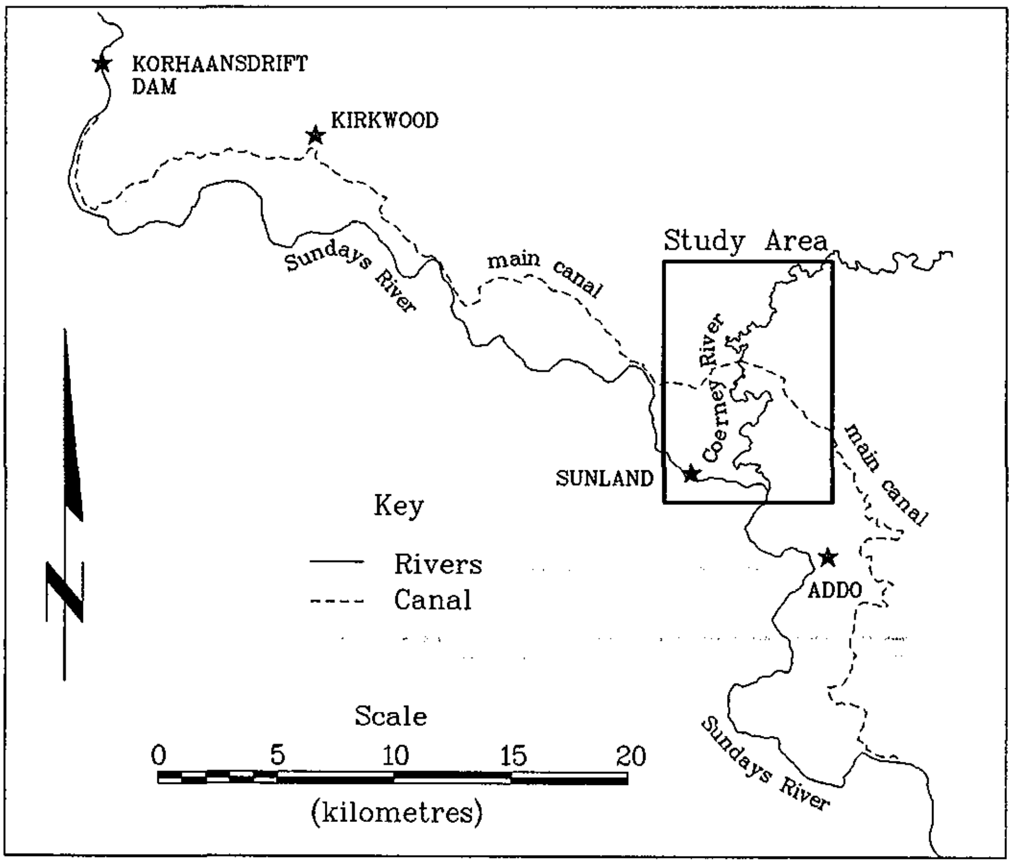


Figure 2.1 Location of the study area in the lower Coerney River valley.

2.1 CLIMATE

Mean annual rainfall for the lower Coerney River valley is approximately 400 mm, which occurs year round but with some increase during the spring and autumn months. High temperatures are experienced in the valley with mean daily summer temperatures ranging from 15° C to 26° C and mean daily winter temperatures ranging from 7° C to 19° C (Schulze, 1965).

2.2 GEOLOGY

The geology of the study area is dominated by recent alluvial deposits, and mudstones and shales of the Kirkwood and Sundays River formations. To gain an understanding of their development it is useful to examine the geology of the area at a regional scale (figure 2.2). The Zuurberg, which is made up of rocks belonging to the Cape Supergroup, is exposed in the north as an outcrop of east-west trending fold ridges, namely the Cape Fold Belt, and is visible from the study area. The tectonic conditions which formed this fold belt between 330-450 Ma continued until the Mid-Jurassic (approximately 215 Ma), after which conditions changed radically (Truswell, 1977). Regional stress patterns generated by shearing along the southern continental margin during the early Cretaceous (65 Ma) resulted in tensional displacements on the southward-dipping faults in the Cape Fold Belt (Tankard et al., 1982). This led to the development of numerous small fault controlled basins. The largest of these is the Algoa Basin which is approximately 4 000 km^2 in extent. It comprises a series of half-grabens which are further dislocated by normal faults, examples of these being the Coega, Zuurberg and Commando faults.

In the Algoa Basin, which is not a simple trough but a complex assemblage of sub-basins, three formations, all of a classic nature have been formally defined (figure 2.2). These formations are the Enon Conglomerate, the Kirkwood, and the Sundays River, together they constitute the Uitenhage Group (S.A.C.S., 1980).

The Enon Conglomerate consists of coarse, poorly sorted conglomerates interbedded with subordinate sandstones and mudstones. This formation represents braided alluvial fan deposits which grade laterally and vertically through fluvial, estuarine and then fossiliferous marine strata (Trustwell, 1977).

A detailed account of the Kirkwood and Sundays River Formations is given by Shone (1976). The Kirkwood Formation is the older of the two formations and has a transitional, often interfingering contact with the Enon (Tankard et al., 1982). It underlies the alluvium upstream from Dunbrody, outcropping to the north and south where it is drained by the tributary Witte and Bezuidenhouts rivers. This formation is composed of red to greenish grey shales and siltstones and fairly massive sandstones, all of fluviatile origin. Some of the gravels contain Dwyka tillite pebbles. Many of the sandstones have a high porosity and permeability.

In the area below Dunbrody, the Kirkwood Formation is overlain by the thinly bedded mudstones and siltstones of the Sundays River Formation. The characteristic colours of these sediments are grey to olive green. It can be inferred from the development of thick bioturbated and rooted red and green mudstones that this formation accumulated in estuarine

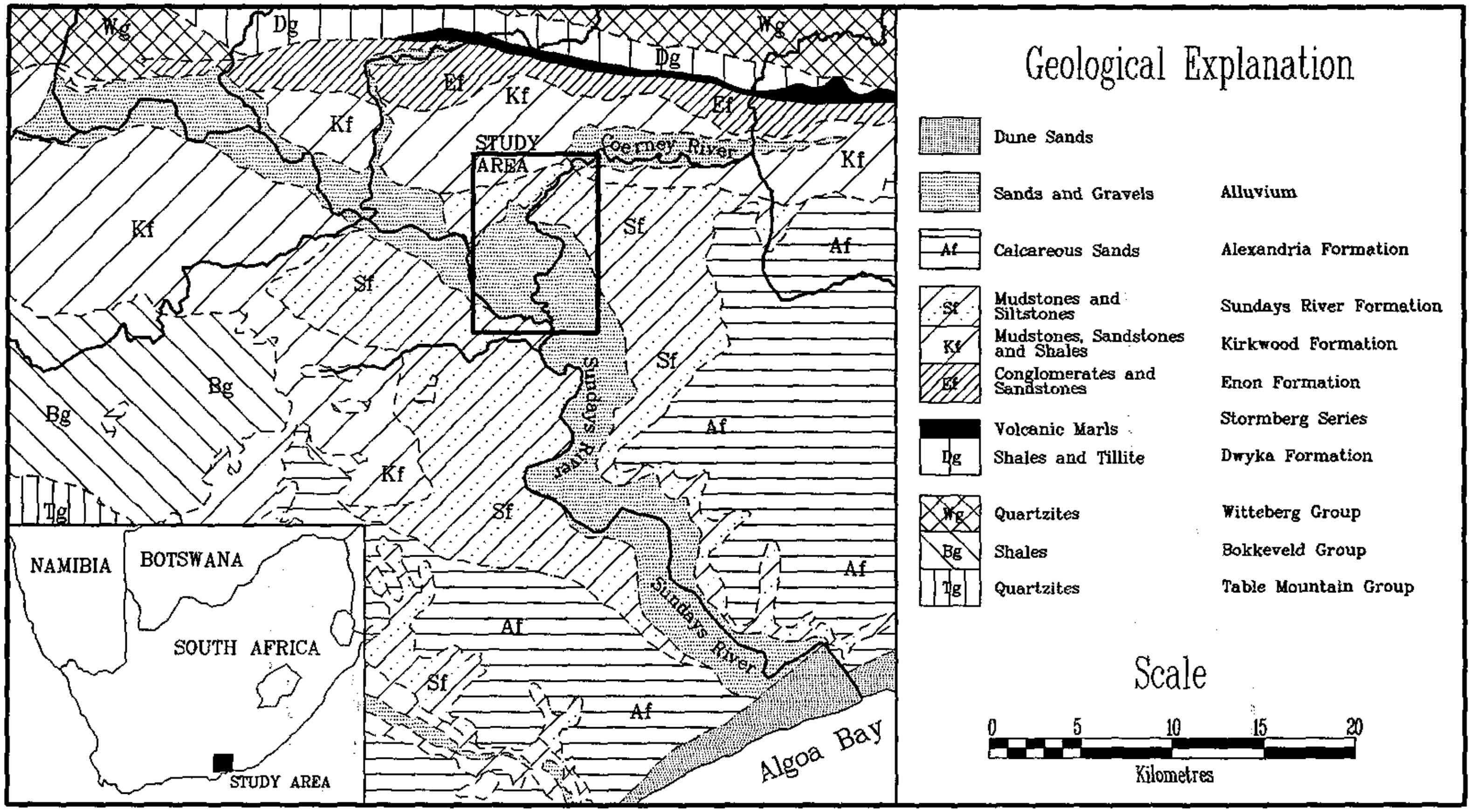


Figure 2.2 Geology of the lower Sundays River valley.

and interdistributary bay environments (Tankard et el., 1982). This being supported by an abundance of marine fossils (S.A.C.S., 1980).

Finally, recent alluvium deposits overlie the previously mentioned formations. These deposits dominate the geology of the lower Coerney River valley and consist of mainly semi-consolidated, interbedded alluvial clays, sands and gravels (figure 2.3). In the Coerney valley these deposits mostly overlie the Sundays River formation, with the Kirkwood formation only exposed to the north of the study area.

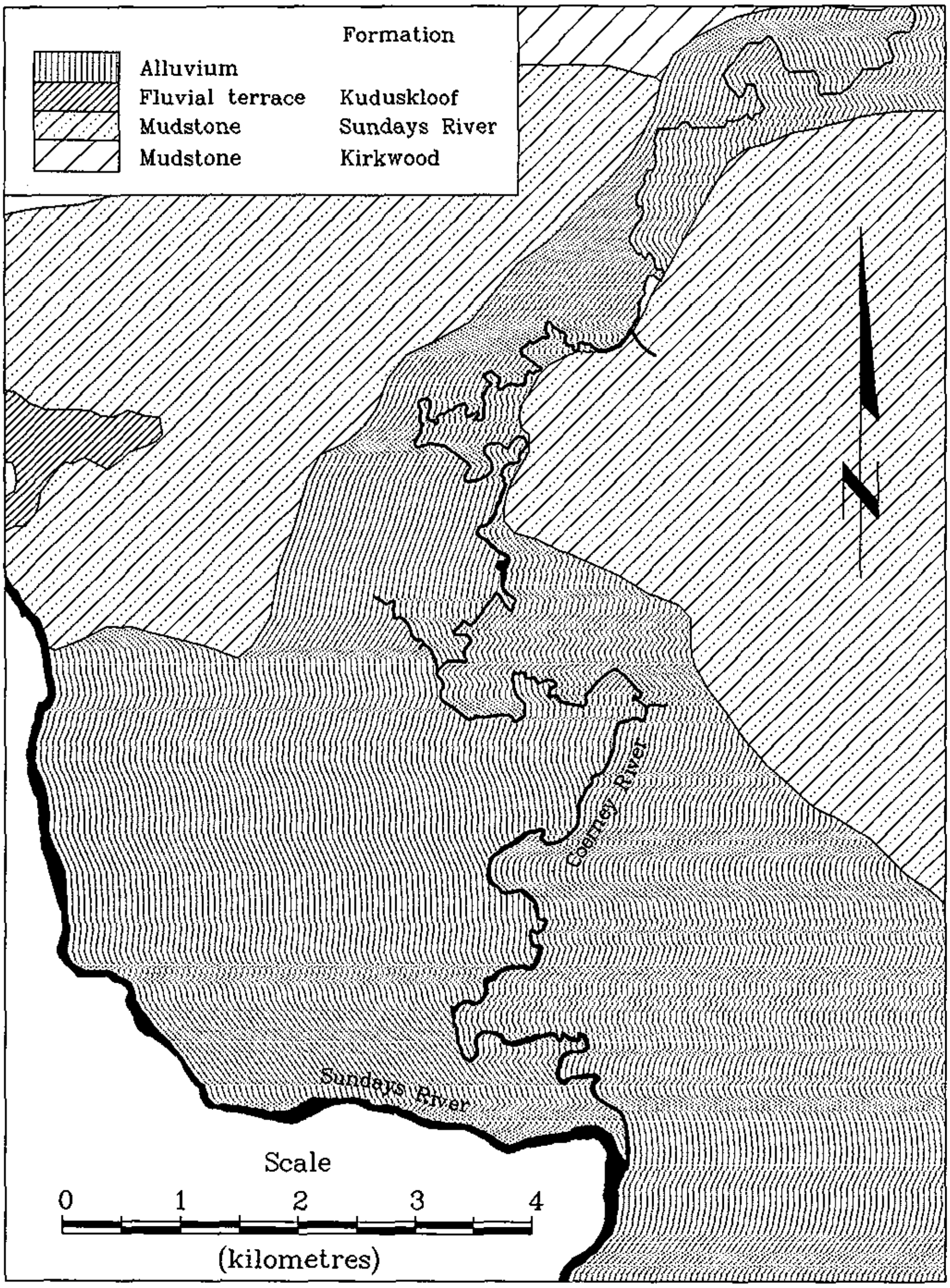


Figure 2.3 The geology of the lower Coerney valley.

2.3 GEOMORPHOLOGY

The geomorphology of the lower Sundays River area is described by Ruddock (1947). Between the scarp and the river a series of high terraces are cut into the Cretaceous rocks. These terraces are dissected by the Coega-kamma, Bezuidenhouts, Kudus and Coerney rivers. A further series of terraces, the lower terraces, are formed in the river alluvium. According to Ruddock (1947) the two sets of terraces are clearly demarcated in the field by a steep and prominent escarpment which is approximately 65m high. Ruddock (1947) recognised four

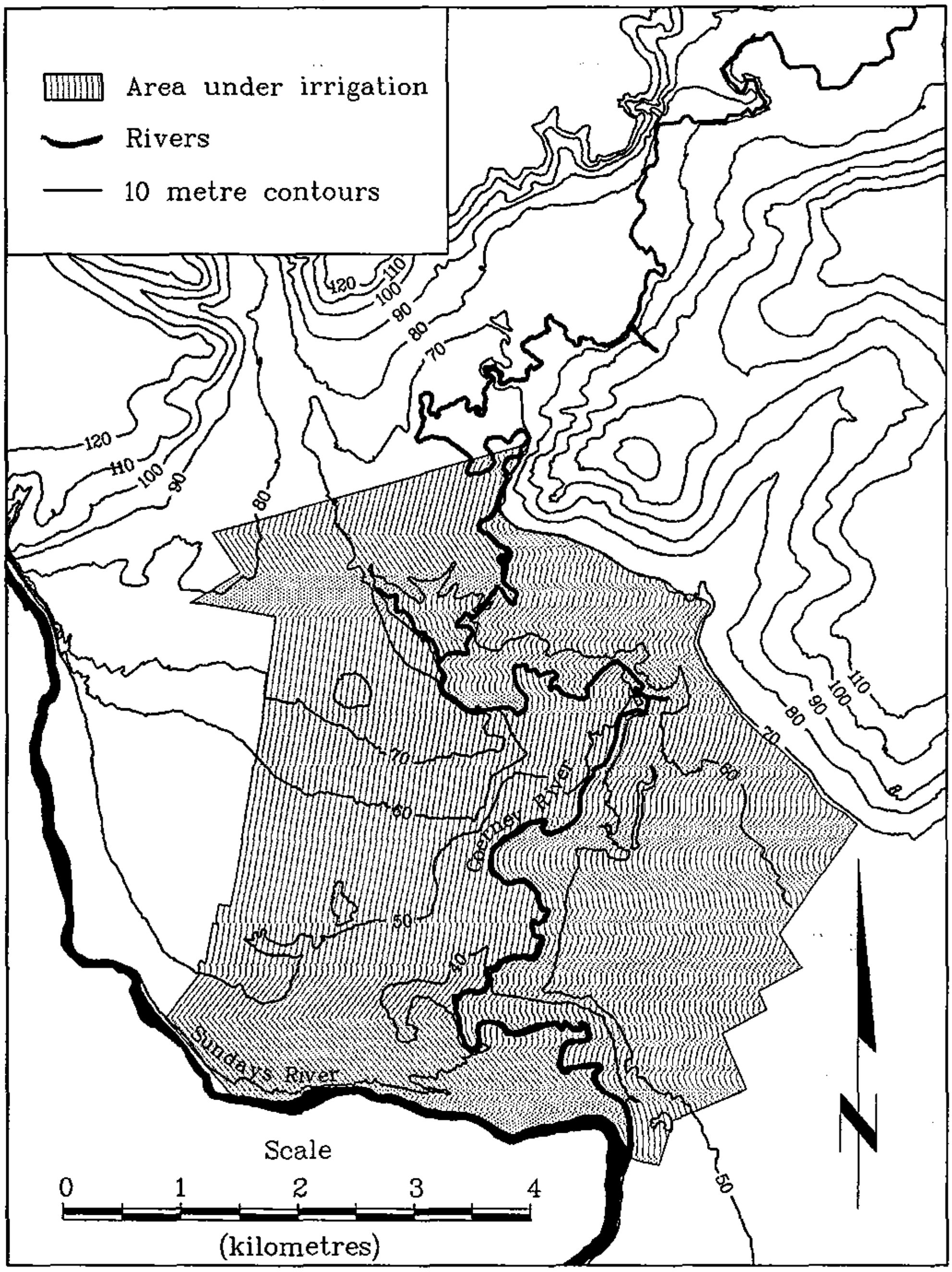


Figure 2.4 Topography of the lower Coerney valley displayed as 10 metre contours.

terraces in the area, which in descending elevation are the Kirkwood, Harveyton, Addo and Cholchester Terraces. A common feature described for these terraces is a significant depth of alluvium overlying a boulder bed. This boulder bed is thought by Ruddock (1947) to represent the former stream bed which was covered by the finer alluvium as the river migrated laterally.

The alluvial terraces in the lower Coerney River are shown in figure 2.5. The Kirkwood and Harveyton terraces can be clearly identified and tend to grade into one another. Above the

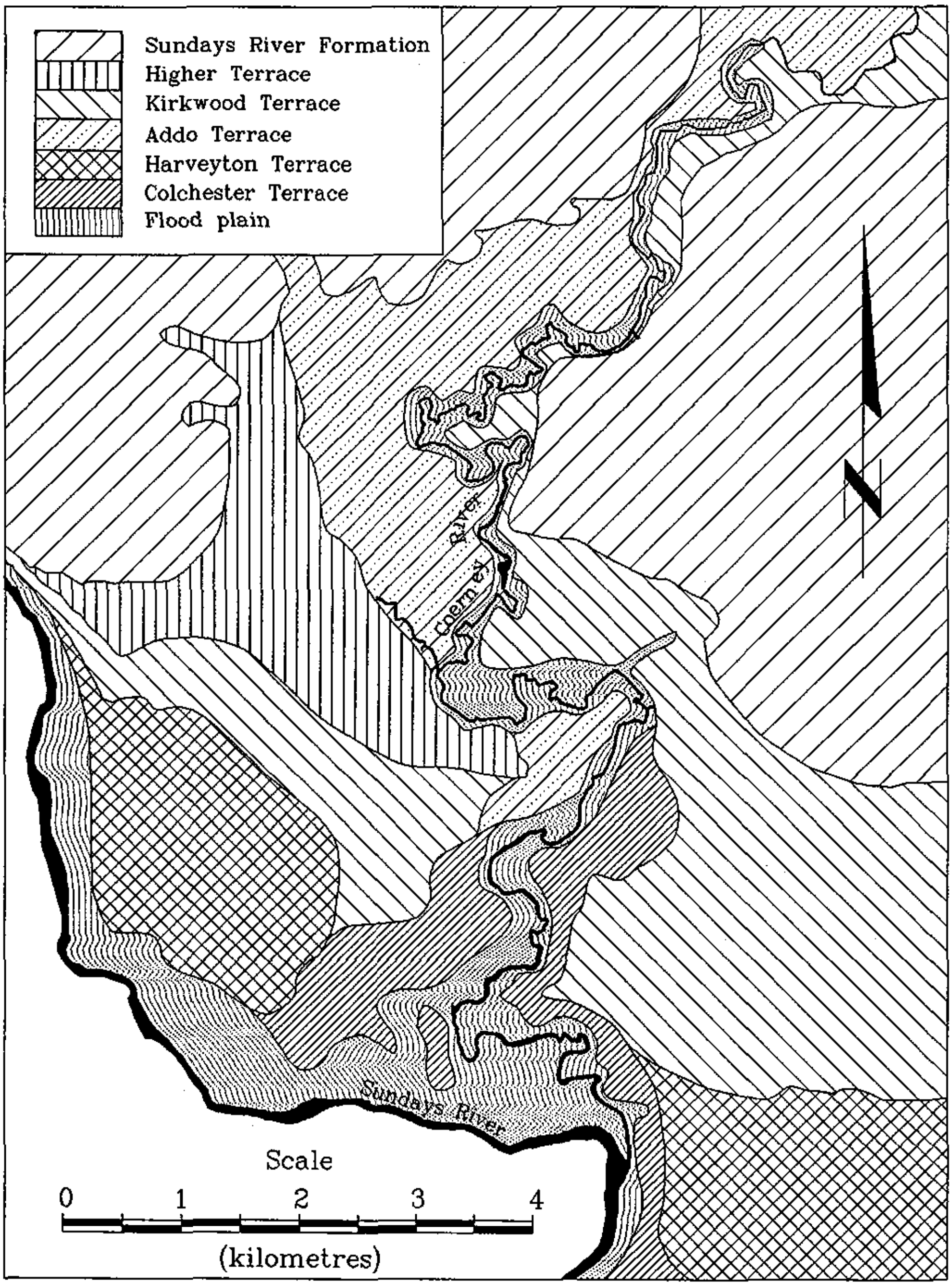


Figure 2.5 Alluvial terraces within the lower Coerney valley.

Kirkwood terrace a higher level can be distinguished. Although of alluvial origin, it may be classified as the lowest member of Ruddock's Higher terraces. The Coerney River cut through these upper terraces which stand up clearly on either side of the river. Associated with the course of the Coerney River three further levels can be distinguished, namely the Addo and Colchester terraces and the flood plain. However, their correlation with their counterparts as mapped by Ruddock for the Sundays River itself is not certain. As shown in figures 2.4 and 2.5 the Addo terrace forms an extensive level surface in the upper section of the basin. Below this area the flow of the Coerney becomes confined between the Higher terrace to the south and the Kirkwood terrace to the north. The Addo and Colchester terraces have been mapped below this constriction.

The Coerney river is incised into the flood plain to a depth of several metres. According to Shone (1965) the present bed of the river consists of a rubble layer, well mixed with silty material. The extent to which the river overtops its banks and inundates the flood plain is unknown. In the upper valley the flood plain is noted to be a narrow feature, more or less confining the meandering course of the river. As the river approaches its confluence with the Sundays River the flood plain feature becomes wider and more complex.

2.4 SOILS

Within the Coerney River valley Shone (1965) distinguished four main soil types. First, silty soils of the Coerney River alluvium occur adjacent to the river. Secondly, on the lower plateau soils have developed from coarse textured sandy alluvium and are covered by a fine textured surface aeolian deposit. The distribution of these soils is largely coincidental with the Colchester terrace as mapped on figure 2.5 and the Addo terrace of the upper valley. For the northern part of the study area Shone (1965) describes similar soils, except that the sandy alluvium overlies a 3 m rubble layer of coarse sand, pebbles and boulders. Again the top soil is of fine textured aeolian origin. The fourth group described are colluvial soils derived from the Sundays River Formation. These soils occur along the outer margins of the Kirkwood Formation. A fine textured colluvial top soil overlies the sandier alluvium of the Kirkwood terrace. The most recent soil mapping in the study area was undertaken by the Department of Agriculture in 1979 (S.I.R.I., 1979). The majority of the soils in the Coerney River valley are mapped as Oakleaf form with small areas of Valsrivier which correspond to Shone's colluvial soils (figure 2.6).

The Oakleaf form is developed on unconsolidated material (in this case alluvium) with an orthic A horizon overlying a neocutanic B horizon. All the Oakleaf soils of the Coerney area belong to the calcareous group and have an apedal to weak block structure. The mapping units are distinguished on the basis of colour and texture of the B horizon. A comparison of figure 2.5 and 2.6 shows that the soil units tend to be related to the terraces.

Soils on the flood plain and the Colchester terrace are classified as Limpopo series with non-red B horizons and fine sandy clay loam textures (Oa2). No soil unit is mapped corresponding to the silty alluvium described by Shone (1965). Soils on the Addo terrace are also mapped as Limpopo series but can be distinguished from the lower terrace soils by their fine clay to sandy clay texture (Oa3).

The Oakleaf soils developed on the Kirkwood and Harveyton terraces as well as the Higher terrace belong to the Letaba/Shigalo series. These soils have a red B horizon. The texture of the B horizon varies from a fine sandy loam (Oa4) to a sandy clay (Oa5). The clay content appears to increase on the Higher terrace and on the more marginal areas of the Kirkwood terrace.

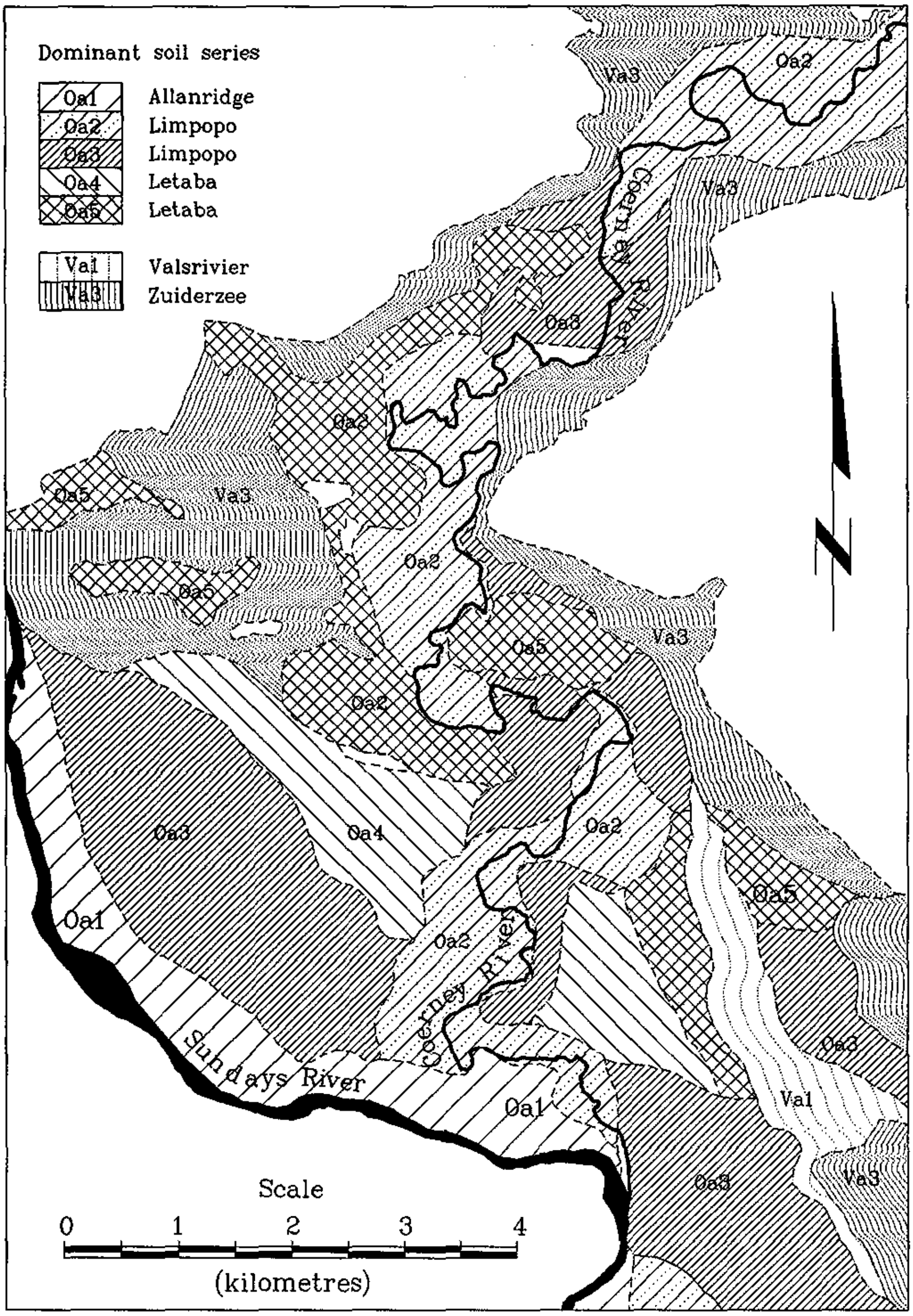


Figure 2.6 Dominant soil series of the lower Coerney valley

The Valsrivier form is found on the contact between the alluvium and the higher outcrops of the Sundays River formation and are derived from colluvial material overlying unconsolidated alluvium. This form is characterised by an orthic A horizon and a pedocutanic B horizon overlying unconsolidated material. The red B horizon which is calcareous has a clay content of between 15 and 35 percent (Mac Vicar et al., 1977). The soils in this area are described as having a moderate to strong blocky structure and a fine sandy clay loam to sandy clay texture.The depth of the B horizon is less than 450 mm.

The soil classification of the Department of Agricultural Development is largely based on the properties of the soil developed on the alluvium and is not directly concerned with the alluvial profile itself. Information on the alluvium has been gained from the logs of boreholes drilled by the Department of Water Affairs as part of the ground water investigation of this project.

2.5 LAND COVER

Acocks (1988) describes the indigenous vegetation of the lower Coerney River as *Valley Bushveld*. However, within the study area their are only small remnants of this vegetation as most areas have been cleared for agricultural purposes. The dominant land cover within the lower Coerney valley is irrigated citrus although large areas above the old canal are still used for live stock grazing, particularly cattle and goats. Within the irrigated area there are also large areas under lucerne and vegetable farming. With the completion of a new higher level canal it is expected that the area under irrigation will increase from approximately 792 ha to over 1 700 ha. Citrus will remain the dominant crop although the proportion of the area used for vegetable farming is expected to increase as farmers attempt to gain returns, on a more short term basis, to offset their more long term and capital intensive citrus developments. There may also be a trend for some farmers to diversify more permanently into vegetables to ensure some income from the local market which they perceive as being more stable and of greater potential.

Traditionally a flood irrigation system was used within the lower Sundays River. However with the development of new orchards and the replacement of old orchards this method is being replaced by the more modern micro-jet irrigation systems. The Coerney area having been more recently developed is predominantly micro-jet, a trend that will continue with the further development of the newly irrigable lands below the new canal.

3. THEORETICAL BACKGROUND

The central aims of this research are to build on the conceptual understanding of the hydrosalinity processes within the lower Sundays River area and to evaluate a number of hydrosalinity models or components of models applicable to irrigation management. However, before undertaking a detailed examination of the field area an overview of the relevant hydrochemical and hydrological processes is presented. This helps justify the research aims and place the overall study into a proper context. As it is important to understand the medium within which the examined hydrosalinity processes operate this overview initially examines the structure and development of alluvial deposits. This is followed by a brief review of the processes and approaches to modelling soil moisture and solute movement.

3.1 ALLUVIAL DEPOSITS AND THEIR FORMATION

It is the alluvial soils of the lower Sundays River valley that are used for irrigation. It is also within the alluvium, perched above the aquiclude of the cretaceous mudstones, that the irrigation return flow concentrates within the local ground water to sustain the base flow of the surface rivers. Therefore, before examining the impact of irrigation on streamflow within the lower Sundays River valley, it is important to gain a conceptual understanding of the formation and depositional characteristics of alluvial deposits.

Alluvium refers to unconsolidated material deposited by fluvial processes on the valley slopes, river banks, flood plain and delta of a river system. The mobilisation of alluvium through the river system occurs progressively in the downstream direction, from the upper catchment towards the river outlet. Erosion and deposition of material may occur almost simultaneously or through a series of temporary storage areas, from which sediment is periodically removed, reworked and redeposited under different fluvial conditions resulting in complex deposits of alluvial material characterised by heterogeneity.

Where there is an abrupt change in slope, these depositional features often take the form of alluvial fans. An alluvial fan is a body of fluvial deposits whose surface approximates a segment of a cone that radiates down slope from the point where the stream leaves a mountain area (Petts and Foster, 1985). Deposition is caused by changes in the hydraulic regime resulting from the sudden reduction in channel slope and the resultant decrease in stream energy which forces the river to deposit some of its load. This is the likely process that lead to the initial infilling of the lower Sundays River valley and its tributaries.

Considerable post depositional modification of valley-fill sediments can take place during migration of the river channel across the flood plain. The valley fill stratigraphy reflects the variability of the dual processes of erosion and deposition of sediment, which have occurred during the formation of the valley fill. Data from most alluvial fills describe distinct stratigraphic units separated by erosional surfaces, differences in sedimentology or lithology and soil development suggesting that fluvial geomorphic activity is episodic. Thus stratigraphic cross-sections of alluvial sequences may be used to produce a chronology of valley-fill development, which incorporates periods of aggradation, erosion and stability. However, as the geometry of a sedimentary unit is not only a function of its original deposition, but is also determined by post-depositional erosion, interpretation of past stream behaviour from stratigraphic evidence is often difficult (Butzer, 1980). One of the main problems is that alluvial deposits record only a fraction of an alluvial cycle and interpretation can be problematic in the absence of suitable dating materials.

On the floodplain gradients are gentler and sediments are generally finer-grained than those comprising valley fill or alluvial fan material. Channels are not confined within narrow valley walls, which means the potential for deposition and lateral channel migrations is increased. Lateral migration of the river channel results in the erosion of existing deposits and the creation of new ones. This reworking of the floodplain alluvium by lateral migration of the stream forms a floodplain composed of lateral accretion sediments, effectively destroying the vertical stratigraphic evidence of the floodplain's origin. Therefore the flood plain is composed of plan segments of different ages which reflect the pattern of historic and recent channel migration and avulsion. The floodplain surface may have a network of channels linking the main river with oxbow lakes and other depressions. Floodplains are thus composed of channel and overbank deposits.

When rivers incise their floodplains the abandoned surface, which is no longer active flooded, becomes a terrace (Schumm, et. al., 1987). The alluvial terrace is considered to represent two successive periods of time: the relatively flat terrace surface develops during a period of geologic and fluvial stability when base level is constant and downcutting is at a minimum, followed by a phase of incision. Terraces border most valleys, and where the valley has been filled with alluvium, it is logical to assume that buried terraces may lie underneath the alluvial floor of the valley. The terrace sequence that forms in the valley-fill alluvium invariably reflects a complicated erosional and depositional history, which can be further confused by episodic erosion and sedimentation (Schumm, 1977).

Traditionally each alluvial terrace has been interpreted as the result of a climatic change, which is not necessarily the causative factor (Schumm, 1973). Channel degradation and terrace formation have been reported to occur simultaneously with aggradation and terrace burial within the same drainage basin (Haible, 1980). So terrace formation does not preclude simultaneous deposition within the same river, and more than one terrace may be formed during a single period of base level lowering.

3.2 SOIL WATER

3.2.1 Soil water movement

Water and air are found in the interstitial spaces between soil particles. The supply of water, its rate of movement and the availability of oxygen are all determined by the number and size of soil pores (Steila and Pond, 1989). However, before rain or irrigation water can enter the soil pores, it must infiltrate the soil surface. Infiltration is the process whereby water enters the soil through surface void spaces and then percolates downward into the soil. The amount of water that a soil can absorb within a period of time is its infiltration capacity, which is unique to each soil. Many factors influence the infiltration capacity, resulting in large temporal and spatial variations in this soil characteristic.

The antecedent conditions of the soil influence both the capacity and rate of infiltration. If the soil is dry before infiltration begins, wetting the surface generates a strong capillary potential immediately below the soil surface which supplements the gravitational force to increase infiltration (Steila and Pond, 1989). If the soil is already wet, the amount of water that can be absorbed by the soil is reduced which inhibits infiltration. Under these conditions, ponding of water will occur on the soil surface, and eventually surface runoff will result. Clay minerals occupy less space when dry than wet, leaving cracks within the soil. If the soil is initially dry, infiltrating water will move rapidly under macropore flow. If the soil is initially wet, or as infiltration proceeds, water is absorbed by these minerals which begin to swell and block the macropores. Thus flow rates will decrease rapidly if swelling clay minerals are present.

Raindrop impact and agricultural activities such as traffic and livestock trampling can compress soil particles and break down soil aggregates. This increases surface bulk density, thus reducing porosity and hence reducing infiltration rates. Interstitial spaces may also become blocked by the inwash of fine particles released from the aggregates, reducing permeability and further inhibiting infiltration.

The rate with which water flows through a soil, is the soil permeability. Pore size and interconnectivity determine the permeability, which is a measure of the effectiveness of the links between the pores. A soil may have high porosity but poor permeability if the pores are not connected. Over time, preferential flow through slightly larger pores will lead to a gradual enlarging of the pore and its links to the next pore. This will further encourage preferential flow, so rapid flow may occur in one zone while slow or non-existent flow may dominate in another area. This heterogeneity of flow may become entrenched in the profile over time, resulting in very different behaviour and properties over a short distance within the same soil. Together, the porosity and permeability determine the infiltration rate of incoming water. As both usually decrease with depth, the rate of infiltration will also decrease.

3.2.2 Soil water retention

Once water has entered the soil profile it is acted on by two main forces. Capillary forces act in all directions to hold a layer of water around each soil particle. Gravitational forces act downwards to draw the water out of the profile into the receiving groundwater. The water film immediately in contact with the soil particle is rigidly held by adhesive electrostatic forces between the dipolar water molecule and the electrically charged soil surfaces. These molecular forces are so strong that this water is considered to be part of the particle crystal lattice, and exhibits little or no movement.

Further out from the soil particle surface, but still within the micropore space, a liquid envelope of water is held in place by capillary forces to a thickness of approximately fifteen molecular layers (Steila and Pond, 1989). These forces are created by surface tension and act equally in all directions to hold a layer of water around soil particles. Capillary forces exceed the gravitational force until a critical pore radius (r) is reached. The capillary forces are then overcome by gravity and the water drains from the soil pore. The strength of the capillary force (Φ) holding this water is determined by:

$$\Phi = (2\yen\cos\tau)/g(q_l - q_g)R \quad (3.1)$$

where $\yen$ = surface tension of water
τ = contact angle of water
q_l = density of the liquid
q_g = density of gas
R = pore radius
(after Hillel, 1980)

Larger pores have greater menisci, so the force with which water is retained is less than for smaller pores. Overlying the water table is a layer of soil at capillary saturation where capillary forces dominate. In an upward direction there is a systematic decrease in soil water as water films separate into discrete menisci. The depth of this capillary fringe depends on the soil pore size. Within a clay soil the capillary fringe is greater than that for a sandy soil. Capillary forces exert an attractive force on water within the soil, which is measured as the matric potential of the soil. When both macro- and micropores are saturated, the soil particles are surrounded by the maximum thickness of water and the matric potential is zero. At this point the capillary forces are equal to zero, and any increase in soil water will now drain under the influence of gravity. Conversely as the soil dries out the water is held with increasing force, as would be expected according to equation 3.1.

Gravity exerts a constant downward force on the soil water and leads to drainage out of the soil, imparting a uni-directional vector to soil water flow. Water affected by gravity is usually contained in macropores which have low capillary potential as indicated by equation 3.1. Such water is only temporarily held within the soil and drains rapidly, so is of limited use for extraction by plants.

Plants obtain most of their water from capillary water. As the soil dries out the matric potential increases to approximately -15 MPa at which point plants can no longer extract water from around the soil particle (Lesk, 1982). This tension is termed wilting point, and if the soil dries out further, permanent wilting point is reached, from which plants can no longer recover. Once all gravitational water is drained and the maximum amount of water is retained in the soil, the soil water content is at field capacity. This is dependent on the number of small pores and makes up on average 30% by volume for a clay soil, and less than 10% for sandy soils (Foth, 1978). Plant available water is the difference between the field capacity and the wilting point of plants. Fine particles increase the field capacity and thus also the available water content.

3.2.3 Variability in Soil Hydraulic Properties

Considerable differences in hydraulic properties of soils can occur with spatial location, even within a given soil type. Variability occurs on a wide range of scales, from micro- to field-wide fluctuations, and in three dimensions. Lateral variation in hydraulic properties, such as hydraulic conductivity, is not as great as variation in the vertical direction in most soils, which are anisotropic in all directions, especially vertically (Mantaglou and Gelhar, 1987). Temporal changes can be even greater than spatial variations.

Variations in soil properties are usually not completely random, but have structural arrangements. Changes over time are also generally ordered, in seasonal cycles or showing net directional progression. Capillary forces, mean soil water content and soil water capacity show large scale hysteresis in stratified soils (Mantaglou and Gelhar, 1987). This is due to the spatial variability of the local hydraulic soil properties, rather than to hysteresis in the parameters themselves.

3.3 MATHEMATICAL DESCRIPTIONS OF SOIL WATER MOVEMENT

An understanding of the factors controlling soil moisture and its movement facilitates the mathematical or modelled representation of these phenomena. However, as soils are highly variable, both spatially and temporally, attempts to quantify the soil system mathematically usually involve some degree of simplification. Initially, the easiest assumption to make is that the soil is completely saturated.

3.3.1 Saturated Conditions

The water balance of the soil can be used to account for the incoming and outgoing fluxes of a soil compartment. Under saturated conditions where most pores are considered filled, and the hydraulic conductivity constant at maximum rate, the difference between incoming moisture and stored water equals the outgoing water (equation 3.2). Under saturated conditions:

$$\delta S = I + Qu - E_t \qquad (3.2)$$

where δS = change in water storage
I = infiltration
Qu = net upward flow
E_t = evapotranspiration

The application of Equation 3.2 is difficult due to problems in quantifying the capillary rise (Qu). An alternative approach is described by Darcy's Law, which gives the flow rate (Q) as the product of the matric potential gradient and the hydraulic conductivity of the soil over a period of time. Darcy's Law does not require direct measurement of upward flux as it assumes that a positive change in matric potential indicates wetting conditions, hence downward flux, while a negative change in potential indicates drying (Equation 3.3):

$$q = -Ks.H \qquad (3.3)$$

where q = moisture flux
Ks = saturated hydraulic conductivity
H = hydraulic gradient = $(\delta(g + c)/\delta z)$
where g = gravity forces
c = capillary forces
z = depth

3.3.2 Unsaturated Conditions

Darcy's Law can also be applied to unsaturated conditions if a steady state is assumed (Richards, 1931) and the unsaturated hydraulic conductivity is measured. In a partially saturated soil the effective volume in which flow can occur is reduced as the larger pore spaces are filled with air. If these empty pores were filled with impermeable solid material, the conditions of flow would be analogous to those found in saturated media with a reduced hydraulic conductivity. The transition from wet to dry conditions is characterised by a large decrease in hydraulic conductivity which occurs over several orders of magnitude (Hillel, 1980; Remson, et. al., 1971). Assuming steady state:

$$q = -Ku.H \qquad (3.4)$$

where q = moisture flux
Ku = unsaturated hydraulic conductivity
H = hydraulic gradient

Capillary forces act in all directions, while gravitational forces (z) act in the vertical direction only. Equation (3.4) may now be expressed in three dimensions (x, y and z):

$$q = -Ku[\delta(g + c)/\delta z + \delta c/\delta y + \delta c/\delta x] \qquad (3.5)$$

The steady state assumption can be avoided if equation (3.5) is combined with the continuity equation (equation 3.2), hence accounting for changes in water content with time:

$$D\theta = \delta/\delta z[Ku(\delta c + 1)/\delta z] \qquad \text{vertical direction}$$

$$D\theta = \delta/\delta x[Ku(\delta c)/\delta x] \qquad \text{horizontal direction}$$

where $D\theta$ = $\delta\theta/\delta t$ = change in soil water content with time
θ = soil water content

This form of Darcy's Law, applicable to unsaturated conditions, is known as Richards Equation. Unsaturated conductivity varies with water content, and Richards Equation assumes that conductivity at a given water content is equal in all directions. In field soils, which are largely stratified, layering results in a variable conductivity in the vertical direction, which may also influence movement in the horizontal direction as well. If the wetting front reaches a dry, coarse layer, downward flow is reduced as the hydraulic conductivity of the dry soil will be low. Further infiltration will then depend on the magnitude of the hydraulic gradient, with water movement being restricted until the gradient can overcome the forces of water retention in the finer grained layer. A dry fine-grained layer would wet rapidly due to its high matric potential, but would have a smaller hydraulic conductivity and so the speed of infiltration would be reduced. This would also lead to saturation above the impeding layer. This means stratification acts to enhance lateral flow over flow in the downward direction, thus reducing the applicability of Richards equation to the prediction of soil water movement.

3.4 SOIL WATER CHEMISTRY

All nutrients required by plants other than the gases O_2 and CO_2 are taken up by roots as solutes from the soil solution. The concentration of chemicals within the root zone is controlled by a number of factors including the soil mineralogy which releases ions during the weathering process. Salt inputs through irrigation and rainwater provide further chemical enrichment of the soil solution and climatic factors such as evaporation may concentrate salts in the upper soil layers.

Solutes and nutrients are conveyed to the roots through transport processes, which are also responsible for nutrient loss through drainage and leaching. Opposing the transport processes are storage mechanisms which can retain plant nutrients against drainage and leaching. These storage processes are the mechanisms of cation and anion adsorption and exchange. Clay platelets and some organic particles carry a net negative charge on their flat surfaces. There is a tendency for these particles to form a chemical bond with positively charged cations that are carried in the soil solution. In this way an ionically balanced particle is formed. The charge on the particle determines the number of positive ions that can be attached to its surface. As the soil solution is continually carrying new cations through the soil, cations on a clay surface may be replaced by other cations with a greater affinity for the clay particle, depending on relative ionic concentrations and rates of ionic reactions. Clay sites where cations may be attached are known as exchangeable sites, and the number of cations that can be attached in this way is termed the cation exchange capacity (CEC) of a soil. The CEC of a soil is important in controlling the solute content of soil water. Indirectly it also influences the physical structure of the soil. If all exchangeable sites of a clay particle are filled and it has a neutral charge, it cannot form an aggregate with other clay particles. The porosity, permeability and rate of infiltration of the soil are therefore reduced.

Micropore water is in close contact with the soil particle surface as it is held in place by molecular and capillary forces. At such small distances, chemical reactions between ions held electrostatically on soil particle surfaces can occur in response to concentration gradients within solution. This interaction takes place slowly and over molecular distances, which means the immediate soil solution may be in chemical equilibrium with the particle surface, but in disequilibrium with more distant soil solution. Once cation exchange reactions occur, the released ions move by diffusion in response to the concentration gradient that exists within the laminar zone of the micropores (Ross, 1989). These solutes reach larger pores where soil water is less strongly held, and flow is more rapid. The micropore water chemistry therefore, is dictated by the mineralogy of the soil particles around which it is held.

Macropore water moves primarily under gravitational forces, and hence moves more quickly through the soil profile. This rapid movement does not allow the macropore flow to come to a chemical equilibrium with the soil particles, which are largely bypassed. The chemistry of macropore flow therefore is determined by the chemistry of the infiltrating water, irrigation or rainfall. The concentration of non-reacting solutes is fairly uniform across the macropore cross-section due to mixing during turbulent flow (Ross, 1989).

3.5 SOLUTE MOVEMENT

The movement of specific ions through the soil matrix is difficult to describe or predict accurately, as the processes of precipitation, dissolution, reaction, and exchange among others serve to interfere with solute movement (Ross, 1989). Negative ions such as Cl^- and NO_3^- tend to be non-reactive, resisting interaction with the soil and are usually selected for modelling solute movement.

Chemicals within the soil may be transported both horizontally and vertically, depending on several major physical processes (Lindstrom and Piver, 1986). Solute movement is dependent on soil water and the transport of solute is assumed to be governed by convection, the viscous movement of the soil solution, and dispersion, thermal motion within the soil solution (Bresler et. al. 1982). The transport of solute through the unsaturated zone to the water table is a complex process, the understanding of which is based on laboratory conditions. The extrapolation of laboratory results to actual field conditions is doubtful (Gvirtman et. al., 1988), but most theory is necessarily derived from the controlled conditions possible only in the laboratory. Theories of transport consider the soil as a continuous porous medium, with equations being derived for a representative unit volume which is large enough to express its properties in terms of statistical averages.

Two mechanisms dominate the transport of solutes through the soil. Firstly, water moving through the soil carries solutes with it by means of convective transport or mass flow. Secondly, the dispersion or migration of solutes under osmotic potential from areas of high to low concentrations within soil water occurs continually by means of diffusion.

3.5.1 Convective Transport/Mass Flow

Solute transport is made up of two convective or mass flow components. In larger pores, turbulent flow dominates and fast convection occurs. Slow laminar transport occurs adjacent to particle surfaces and in micropores. Soil water movement occurs when a difference in water potential exists. Flow is determined by pressure gradients, and when changes in soil water occur through infiltration, redistribution and evapotranspiration, the dissolved salts move with the water.

Convective transport depends on the macroscopic flow velocity. Due to the soil porosity, actual porewater velocity is distributed around an average value and is determined by the distribution of pore sizes and pore shapes. Velocity is faster in large pores, and in the centre of the pores. Water and solutes move at different rates through soil and may be acted upon, transformed, or retarded during their movement through the soil. Where conservative solutes are considered no gains or losses and no solute/surface or solute/solute interactions occur (Ross, 1989). Under these conditions, the convective flow of solute associated with water movement is expressed by:

$$Jc = q.C \qquad (3.6)$$

where Jc = convective flux of solute
q = volumetric flow rate
C = concentration of solute

To estimate the travel distance of the solute, the average apparent solute velocity is considered. Actual velocities vary over several orders of magnitude within pores and between pores, so the average is at best an approximation. Solute velocity is assumed to be equal to moisture velocity, which is applicable to non-reactive ions only:

$$v = q/\theta \quad \text{(3.7)}$$

where θ = volumetric soil moisture
v = average solute velocity
q = volumetric flow rate

Transport of solutes seldom occurs by convection alone as solutes move within flowing water in response to concentration gradients by the processes of diffusion and dispersion.

3.5.2 Diffusion

Solute movement is affected by solute concentrations, so chemical behaviour at the molecular scale must be considered. All solute molecules exhibit random motion which results in the net movement of ions from high to low concentration until the solution is uniform. The speed of equalisation depends on the concentration difference, thus solute transport by molecular diffusion depends on the concentration gradient of the ion. The average macroscopic flow rate of solute molecules in a uniform aqueous medium is proportional to the concentration gradient ($\delta c/\delta x$) and to the cross-sectional area. This is Fick's first law for saturated media.

$$Jd = -Dw.\delta C/\delta z \quad \text{(3.8)}$$

where Jd = rate of diffusion
Dw = diffusion coefficient for the solute diffusing in water
C = solute concentration
z = distance in direction of flow

Rewriting the equation for unsaturated conditions:

$$Jd = -Ds.\ \theta.\ \delta C/\delta z \quad \text{(3.9)}$$

where θ = volumetric soil water content
Ds = diffusion coefficient in the soil

Accounting for three dimensional diffusion:

$$Jd = [-Ds.\delta Cx/\delta x] + [-Ds.\delta Cy/\delta y] + [-Ds.\delta Cz/\delta z] \quad \text{(3.10)}$$

In unsaturated conditions the volume of water available for diffusion is reduced, so Ds can increase by several orders of magnitude over Dw. The path of diffusion becomes more tortuous when there is less water available to move through, which further reduces the actual value of Ds. Mahtab, et. al. (1971) found Ds to increase linearly with water content of the soil. As the water content of a soil decreases, the cross sectional area available for diffusion becomes smaller and the ions have to travel a longer distance to reach a given point. Other factors such as viscosity and anion exclusion become more influential as water content decreases.

According to Ross (1989), diffusion coefficients in soil are controlled by:

i. The state of the medium within which diffusion occurs. Diffusion is most rapid in gaseous media, followed by liquids then solids.

ii. Soil water content. A two- to four-fold increase in diffusion of K^+, Ca^{2+} and Mg^{2+} was measured by Schaff and Skogley (1982) when soil water content was increased from 10 to 20%. Table 3.1 shows the effect of increasing soil water on the diffusion coefficients of Na^+, Cl^{-1} and PO_4^{-1}. When soil moisture content increases from 20 to 40%, the coefficients increase by an order of magnitude for Na^+ and PO_4^-, and by 3-fold for the Cl^{-1} anion.

iii. Tortuosity of porespaces. Smaller pores, and increased tortuosity of flow paths within the soil matrix decrease ionic diffusion of all ionic species. This was first recognised by Nye and Tinker (1977), who introduced an impedance factor to modify the diffusion coefficient:

$$Ds = -D.\theta.f.\ \delta C/\delta z \qquad (3.11)$$

where	Ds	=	diffusion coefficient of solute through soil
	D	=	diffusion coefficient of solute in free solution
	θ	=	fraction of the soil volume occupied by solution
	f	=	impedance factor for tortuosity of diffusion pathway
	C	=	concentration of solute in solution
	z	=	depth

(Nye and Tinker, 1971)

Empirical expressions have been derived for diffusion governing the movement of a non-consumptive, non-reactive solute, Phosphorous (Mahtab, et. al., 1971). Ds was governed by the amount of solute in solution and the rate at which solid phase solute is released into solution. The available solute increased as the clay content of the soil increased, and therefore as the clay content increased so did Ds, on average 5 fold in a soil with 5 times as much clay. The Ds values were significantly related to clay content at a 99 % level of confidence. The writers suggest that the increase in Ds values with increased clay content is probably due to the "tortuosity" factor, which also becomes larger with clay content.

Equation 3.10 only describes the steady state diffusion processes. To account for transient processes in which rate of diffusion and concentrations vary with time Equation 3.10 is combined with the continuity equation (Equation 3.2):

$$A.\delta C/\delta t.z = A[Jd + \delta Jd/\delta z.z] - A.Jd \qquad (3.12)$$

where A = cross sectional area of flow
C = solute concentration
t = time
z = distance
Jd = rate of diffusion
$A[Jd + \delta Jd/\delta z.z]$ = solute diffusing in per unit time

The diffusion rate (Jd) is not generally constant as it is dependent on both soil moisture and solute concentrations. However, where diffusion is uniform, equation 3.12 reduces to:

$$\delta C/\delta t = -\delta Jd/\delta z \quad (3.13)$$

Convective flow changes the distribution of solutes relative to one another, and induces the second process determining soil solute movement, hydrodynamic dispersion.

Table 3.1 Diffusion coefficient D for selected ions at given soil water.

Ion	Volumetric water content (%)	D (cm^2/s)
Na^+	40	2.2×10^{-6}
Na^+	20	0.2×10^{-6}
Cl^-	40	9.0×10^{-6}
Cl^-	20	2.4×10^{-6}
PO_4^-	40	3.3×10^{-9}
PO_4^-	20	0.3×10^{-9}

(Rowell et. al., 1967)

3.5.3 Dispersion

Mechanical dispersion is due to local variations in the velocity of flows moving through the soil pore structure. These variations in flow result from water moving faster through wider pores, and also as flow velocities increase near the centre of each pore. This variability of flow causes mixing of solutions, the degree of which depends on the average flow velocity, pore size distribution, degree of saturation and concentration gradients (Hillel, 1980).

When convective velocity is high, such as at near-saturation conditions, hydrodynamic dispersion will exceed diffusion. This means diffusion will be negligible in terms of solute movement. However, during unsaturated conditions hydrodynamic flow ceases and diffusion becomes the dominant mechanism for solute movement. A similar partitioning occurs under variable flow velocities. Molecular diffusion is obscured by much larger dispersion effects at high water velocities. As porewater velocity decreases the dispersion coefficient (Ds) decreases and molecular diffusion becomes more important in solute mixing (Kirda et. al,1973).

Mechanical dispersion (Dm) is similar to Ds and has been found to depend, approximately linearly, on the average flow velocity. It has also been found that the dispersion coefficient is independent of porewater velocity (v), and is proportional to the water content (Smiles and Philip, 1978; Elrick et. al., 1979; Smiles et. al., 1981). The dispersion coefficient (Dm) can be expressed by the following:

$$Dm = D0 + Ev \quad \ldots\ldots (3.14)$$

where D0 = molecular diffusion coefficient
E = dispersivity
v = pore water velocity

Equation 3.14 has been widely accepted as being applicable to field conditions (De Smedt et. al., 1986, Kirda et. al. 1973; Hildebrand and Himmelbau, 1973; Yule and Gardner, 1978; Beese and Wierenga, 1983), although it remains controversial. Dispersivity (E) ranges from 0.1 to 10 mm under saturated laboratory conditions, but under field conditions, values of up to 100m have been reported (Freeze and Cherry, 1979). This means it is necessary to calibrate the field dispersivity of a selected ion before equation 3.14 can be used with confidence.

Under unsaturated conditions both mobile and immobile water phases exist (Van Genuchten and Wierenga, 1977; Gaudet et. al., 1977; Rao et. al., 1980; De Smedt and Wierenga, 1984). Differential rates of matrix flow have been modelled by differentiating soil water into mobile, immobile and stagnant phases. A convective-dispersive transfer of solutes occurs through the simultaneous exchange of solutes between the mobile and immobile phases (De Smedt et. al., 1986). The transport of solutes under these conditions is represented by the following:

For the mobile phase:

$$\delta\theta_m/\delta t.C_m = \delta/\delta z.[\theta_m.D_m.\delta C_m/\delta z] - \delta q/\delta z.C_m - ar.\theta_m[C_m - C_{im}] \quad \ldots\ldots (3.15)$$

For the immobile phase:

$$\delta\theta_{im}/\delta t.C_{im} = ar.\theta_m[C_m - C_{im}] \quad \ldots\ldots (3.16)$$

where θ = soil moisture content
C = solute concentration
t = time
z = depth
D = dispersion coefficient
q = volumetric flow rate
ar = rate coefficient of solute exchange between two phases
m = mobile water phase
im = immobile water phase

After sufficient time, it has been shown that the mixing between the phases is such that concentration distributions can be described by:

$$D \quad = \quad \theta_m.D_m/\theta + [\theta^2_{im}v^2]/\theta_m.ar \qquad (3.17)$$

where the symbols are the same Eq 3.16.

From Equation 3.17 it can be seen that the overall dispersion coefficient (D) is the sum of two processes: dispersion in the mobile zone and dispersion due to solute exchange between the mobile and immobile phases. This suggests that when immobile water is present in unsaturated conditions, the overall dispersion can increase significantly, up to 78 times greater than when only mobile water is considered (De Smedt et. al., 1986).

It is doubtful whether the convection and dispersion equations can model solute transport in soils containing a wide range of pore diameters and porewater velocities (Ross, 1989) irrespective of the adaptations and additional subroutines included. This is because different processes are involved, other than the simple convection and dispersion which are invoked as the causal factors in solute transport.

3.5.4 Effect of Soil Properties on Solute Transport

Soil solute concentration and flux are calculated as a function of time and space using macroscopic quantities which vary in a deterministic manner, and are expressed as partial differential equations. To solve these equations it is usually assumed that their flow parameters are uniform throughout the entire field, however in reality fields are seldom homogeneous.

The most significantly variable soil physical characteristics are water and solute transport parameters measured under field conditions (Jury, et. al., 1986). Solute convective velocities and dispersion coefficients typically show large standard deviations, often exceeding 100% of the field mean (Jury, 1985). Such extreme variability means accurate average values cannot be estimated at field scale. This leads to the assumption that mass transport phenomena in field soils are intrinsically erratic processes for which quantitative characterisation can only be achieved using stochastic approaches at the field scale. Properties of the soil are regarded as continuous functions of the space coordinates, of which the horizontal dimensions are very large compared with the vertical direction. As the spatial variability of field soil characteristics must be expressed in statistical terms, stochastic elements are often used instead of, or in addition to deterministic ones.

Although soil hydraulic properties are highly variable, even within a given small field (Russo and Bresler, 1981b), variations are not completely disordered. Structured variation occurs which often describes a probability distribution of these factors. Using actual field variability data, it is possible to analyse how much inherent spatial variability of the soil hydraulic properties exists, and hence quantify the probability density function (Russo and Bresler, 1981a). Solute movement at field scale distances and times has been found to show variability according to the hydraulic conductivity. This has been successfully described by a single realisation of a stochastic function parameterised by spatial co-ordinates (Sposito, et. al., 1986).

Cameron and Wild (1982) compared the ability of three models to predict chloride leaching in South England, and found the most accurate results were obtained using the convective-dispersion (CD) method of Rose et. al. (1982). Leaching predictions were less accurate for actual rainfall events than regular irrigation applications, and predictions were only possible once estimates of field diffusivity were available. This highlights the problem of applying simple transport models to heterogeneous field conditions. Field measurements have shown that there is considerable spatial and temporal variability in soil water parameters and many attempts to predict field leaching have failed due to the unknown local field variability in soil hydraulic properties (Nielsen, et. al., 1973; Biggar and Nielsen, 1976; Jury et. al., 1976 Van De Pol et. al., 1977).

3.5.5 Miscible Displacement and Breakthrough Curves

If a liquid infiltrating the soil is not soluble in the pre-existing soil water then the process of immiscible displacement will occur, whereby the incoming liquid displaces the soil moisture. Most solutions entering the soil are however mutually soluble and mixing will occur with the chemistry of the incoming solution gradually dominating the soil solution. The change in concentration of the outgoing solution over time can be plotted as a breakthrough curve (BTC).

If the pre-existing water were pushed out of the soil without any mixing with the invading solution, the BTC would show an abrupt change in concentration when all initial water had gone. This would result in piston flow through the soil. As most aqueous solutions are miscible mixing is unavoidable, and the shape of the BTC is variable depending on a number of factors. This provides a useful method for deducing processes occurring between inflow and outflow points. Nielsen and Biggar (1962) first noted that if the area under the breakthrough curve below 1 pore volume is equal to the area above the curve greater than 1 pore volume, then no solute-solid interaction has occurred. This means retention or release of solutes within the soil will be indicated by the shape of the curve (Selim, et. al., 1989). Two factors cause mixing and dispersion, resulting in the outflow curve varying:

i. Increase in the pore size and consequent increase in porewater velocity causes increased dispersion.
ii. Increase in the concentration of the inflow solution has a similar effect to increasing pore size. Solutes move from high to low concentrations, and the greater the concentration gradient, the more rapidly diffusion will proceed.

3.5.6 Anion Exclusion

Soil clay particles and humus surfaces exhibit negative charges which repel anions electrostatically. These anions are then concentrated in the centre of the pores where the velocity of flow is relatively faster, and the water volume available for transport 10-20% less than that available for cations (Wild, 1981). This means that anions move more rapidly than positive ions, and can also exceed the average velocity of the soil water. Smith (1972) observed that Chloride moved 1.04 - 1.67 times faster than the average water velocity. If the net anion flux exceeds the net water flux forward displacement of the BTC will result.

Ross (1989) identifies the following factors affecting anion exclusion:

i. Anion exclusion increases with increasing anion concentration.

ii. Exclusion increases with anion valency. More electronegative ions will be more strongly repelled by the soil particle surfaces, and to a greater distance than less electronegative ions. The following order of repulsion has been reported: $Cl^- = NO_3^- < SO_4^{2-} < Fe(CN)_6^{4-}$ (Mattson, 1929).

iii. Exclusion decreases with soil pH, since this decreases the net negative charge on soil colloids. At low pH, there are more Hydrogen (H^+) ions in solution, which readily occupy the cation exchange sites, rendering the soil particle neutral. The particles will no longer be able to repel anions, hence reducing the anion exclusion effect.

iv. Exclusion decreases with increasing cation saturation of the soil. This is a similar effect to increased pH, except the exchange sites are filled with cations other than H^+.

v. Exclusion increases with increased density of negative charge on the particle surface. This means the particle can exert a greater repulsive force on the anions.

The effects of anion exclusion result in negative ions being rapidly propagated through the soil, usually at rates faster than the net moisture movement. As negative ions such as Cl^-, and NO_3^- are conservative, and do not readily become involved in soil reactions or cation exchange, these are often selected as tracers for soil solute movement (Ross, 1989). However anion exclusion means the movement of anions is not necessarily representative of the net soil moisture movement, or movement of other solutes. The use of anions may lead to over-estimation of movement, conversely the use of cations may lead to underestimation as such ions are readily adsorbed onto exchange sites.

3.5.7 Solute Reaction

Ion adsorption/desorption during solute transport has been incorporated into a number of models (Wagenet, 1984).The problem in incorporating ion exchange into models of transport for reactive solutes is due to the complexity of the soil system and the numerous interactions of competing ions (Ross, 1989). Simple one-dimensional transport models have been developed for solute leaching models using non-reactive ions such as Chloride and Nitrate. Both laboratory and field validation have concentrated overwhelmingly on non-reactive species, but this is inadequate to describe the gross solute distribution within a field soil (Ross, 1989). Therefore there is a clear need for further field validation using a wider range of ionic species, including cations and an examination of the effects of anion exclusion on so-called non-reactive ions.

3.6 HYDROSALINITY MODELLING

Crop production is not solely determined by the quantity of water applied. The timing of water applications and the quality of the water applied are also important. This is especially important within semi-arid environments where there are limited water resources, high evaporative losses and salinity problems are common. Within these environments there is a need for a better understanding of the hydrosalinity processes and improved information for irrigation management. A number of hydrosalinity models, which in theory are ideally suited to meet these needs, have been developed to examine and manage soil water and solutes movement. However, little general use has been made of these models beyond their initial development and testing (Wagenet and Hutson, 1986 and 1987; Russo, 1991). This may in part be due to the intensive data requirements of the more physically based models and also to the differing degrees of success experienced to date.

A major problem is the spatial variability of input parameters required for physically based, deterministic models, even at the field scale. A solution may be the development of stochastic type models, where spatial variability of the soil water and solute transfer properties of the soil can be considered in terms of probability functions (Dagan and Bresler, 1979; Simmons, 1982; Mantoglou and Gelhar, 1987; Tillotson, et. al., 1988). However, even the determination of parameters for stochastic models may require an intensive sampling programme to adequately define statistical variations at the field scale. It is therefore important that the purpose, and therefore the required resolution, of a models output should be clearly examined before model development or selection is undertaken.

Research models are generally unsuitable for management purposes due to their intensive data requirements, and the detailed theoretical knowledge required to apply and interpret their results. Also, they are often less easily transferred from the developer to other users (Wagenet and Hutson, 1986). However, the strength of these research models is that they allow for the testing, in a comprehensive and integrated manner, of current knowledge of the processes controlling water and solute movement. Although a number of these models are potentially applicable to a wide range of applications, few studies are reported in the literature (Wagenet and Hutson, 1986).

Management models are less data intensive and are commensurately less quantitative in their ability to predict solute and water movement under field conditions. Unfortunately, few of these models have been widely tested against field data and little attention has been paid to their potential application for the purposes of managing saline irrigation water (Addiscott and Wagenet, 1985).

This overview of hydrosalinity modelling does not set out to provide an up to date and definitive description and evaluation of the hydrosalinity models currently available. This task has been carried more than adequately by Moolman (1992). However, before evaluating several rootzone hydrosalinity models within this report it is important to define some terminology and to address some commonly experienced problems related to hydrosalinity modelling.

3.6.1 Theoretical Bases for Hydrosalinity Models

Developers of hydrosalinity models have adopted either one of two common approaches, the thermodynamic approach, where water moves according to differences in potential energy, and the capacity approach where a soil layer retains water to a defined upper limit before draining to a deeper layer in the soil profile.

The capacity approach is based on a simple water balance, using relationships that are largely empirical. The maximum volume of water that an individual soil layer can hold is defined as a fixed amount. Infiltrating water moves into the soil until this limit, field capacity, is reached and any excess then moves down into the next layer (Moolman and de Clercq, 1990). Information on the saturated hydraulic conductivity and hydraulic gradients are usually not required. Soil water movement is capacity driven and only flow in the downward direction is normally considered. Crop water uptake is simulated through the inclusion of a simple sink term that accounts for evapotranspiration, although some models distinguish between evaporation and transpiration. The capacity approach assumes steady state soil salinity conditions. Processes included in these models are usually those of dissolution and precipitation in a simple mass balance between incoming and outgoing solute loads for respective soil layers. The assumption of kinetic equilibrium is generally unrealistic as transient conditions, where the soil loses or gains solutes through time, are a more common phenomena.

With the thermodynamic approach, water and solute transport are explained as occurring due to differences of potential energy within the soil. This arises due to differences in matric potential and hydraulic gradients across some spatial distance within the soil. These models do not assume steady state conditions and Richard's Equation is generally used to describe soil moisture movement. However, transient conditions are theoretically more complex to model and require more intensive data. This approach is therefore not as simple or easily applied as the capacity approach.

3.6.2 Solute Modelling

Although limited to conditions of steady-state water flow in homogeneous soils, the analytic solutions of the classical Convective Dispersion Equation (CDE) have been widely used to model chemical transport and transformations within the soil-water system (Van Genuchten and Wagenet, 1989). Solute transport, according to the miscible displacement theory states that the flux of solute is the result of the combined effects of diffusion and convection:

$$J = Jd + Jc \quad \text{.........} \quad (3.18)$$

where J = solute flux
Jd = solute transported by diffusion
Jc = solute transported by convection

Convection is the chemical diffusion of solute in response to concentration gradients existing in the soil solution. Fick's first law of diffusion is used to describe the diffusion process.

$$Jd = -Ds.\delta C/\delta z \qquad (3.19)$$

where Ds = the effective diffusion coefficient of the chemical in the soil
C = solute concentration
z = depth

Dispersion is the physical mixing resulting from variations in water flow velocity within each pore and between pores. This is mechanical dispersion and can also be described by using an adapted version of Fick's law.

$$Jc = [-\theta.Dm.v.\delta C/\delta z] + [v.\theta.C] \qquad (3.20)$$

where Jc = convective flux of solute
Dm = mechanical dispersion coefficient
v = average flow velocity
θ = soil moisture content
C = solute concentration
z = distance

Combining Equations 3.19 and 3.20 gives the convective dispersion equation of solute transport:

$$J = [-\theta.Dv.\theta.\delta C/\delta z] + [q.C] \qquad (3.21)$$

where q = volumetric water flux
D = Dm + Ds

The use of Equation 3.21 to represent solute transport in leaching models is subject to large spatial differences in the relationship between water flux, water content and the apparent diffusion coefficient.

3.6.3 Validation of Models

Model predictions must be compared with measured data to prove that the model output is a realistic representation of field processes. To compare model simulations with measured data the model input parameters must be known and field data to compare with model outputs must also be available. According to McLaughlin (1988):

i. from a technical or scientific point of view a model is validated when it properly describes the physical processes, and

ii. from a regulatory point of view it is validated when the model yields adequate predictions with the implicit goal being to reduce the risk of making inappropriate decisions from the model results.

However, the criteria for validation are not universally accepted. Loague et. al. (1988) suggest that a model is a good representation of reality, and hence valid, if it can be used to predict certain observable phenomenon with an acceptable accuracy and precision. These writers do not define precision or an acceptable level of accuracy. Statistical measurement of the validity of models is desirable, but data sets of sufficient size to provide a representative range of the model output performance are difficult to provide. Statistical testing and graphical representation are used in model validation, but there is no defined procedure or technique that is widely accepted. The level of acceptable inaccuracy will vary with applications, hence one model may be valid for a situation requiring general trends and qualitative information, such as irrigation management or educational purposes, but invalid for pure scientific research.

3.6.4 Sensitivity Analyses and Model Calibration

Model validation is distinguished from calibration, which is the procedure whereby model output is progressively refined using data specific to the site or area in which it is to be applied. Model parameters are adjusted until the predictions generated are acceptable approximations of these measured data. Validation involves comparison of the model with field data with no alterations made to improve the output performance. Data used for the validation must necessarily be different from that used for calibration, or the validation becomes a meaningless exercise (Moolman and de Clercq, 1990).

The process of collecting and compiling the data requirements of a model may be expensive and time consuming. Therefore, before embarking on this exercise it is often useful to embark on a sensitivity analysis in order to determine the relative importance of each input parameter in determining model output. The objective of the sensitivity analysis is to identify those variables having little or no impact on model predictions. These variables can then be measured using less rigorous and more cost effective means.

3.6.5 Limitations of Modelling

A realistic approach to the abilities and achievements of hydrosalinity models must be developed if such models are to be critically applied (Beven, 1989). The major problems related to the application of hydrosalinity models involves the scale of application, and the spatial and temporal variability of model parameters.

The scale of application often determines which models can be usefully applied (Moolman and de Clercq, 1990). Irrigation and other large scale water uses must be modelled at a large scale, which requires low resolution, averaging models rather than the high resolution models which are more applicable to a smaller scale. In general the data requirements and accuracy of prediction declines with simplification in model approach. The physics on which most model equations are based are small scale and derived from homogeneous systems under laboratory conditions. It is assumed that the same small scale equations can be applied at the field scale using the same parameter (Beven, 1986). Studies have shown that the spatial variability of the hydraulic characteristics are structured and display spatial correlations. However, increasing the scale of averaging reduces the derived variance of parameter values. This is because the averaging process is linear, while reality is non-linear; averaging presupposes the model equations are applicable to the field scale while they are derived on a smaller scale; and the random function is assumed stationary, while in reality it varies both spatially and temporally (Beven, 1989).

Spatial variability has been accounted for through the incorporation of stochastic parameters into models, which describe variables that are intrinsically random. The parameters chosen are those which represent a probability density function of a range of measured samples of the variable. Therefore the stochastic parameter must be calibrated for the area on which the model is to be applied. Universal "constants" are highly generalised and unlikely to provide acceptable model results.

Data limitations restrict the application of physically based models to a far greater extent than theoretical considerations (Beven, et. al., 1980). Physically based distributed models are expensive in terms of both computing and data requirements. High resolution models may provide a better approximation of reality than the simpler models, but the time and expense required to measure and quantify all model inputs may render the model prohibitively expensive for routine use in management. Very simple models can provide a useful, if not more accurate representation of the system and the benefits from additional complexity may not be warranted by the corresponding increase in cost. Often decisions can be made efficiently without the use of mathematical models, and the time and money needed to do the modelling could be better spent addressing real life problems. Field data with proper interpretation may be more acceptable.

Models of the physical environment have been largely untested (Tillotson et. al., 1988). Simulation models can only simulate a solution, the results of which are predetermined by the input (Hillel, 1985). When a model has faulty premises or data, there is a danger that it will gain a false aura of respectability simply because it was processed on a computer. There is a need therefore to substantiate or refute the results predicted by models. This can only be done through controlled and detailed experimentation of the system which the model aims the simulate. As information on the actual field conditions is generated "it is becoming increasingly clear that, at present, only approximate prediction of water movement and chemical distributions can be made" (Wagenet, 1988). Furthermore, a prerequisite for the validation of a model is that the necessary model parameters be estimated independently. The intrinsic interdependence of many hydraulic parameters, and the problems in measurement makes this ideal difficult to fulfil.

Despite these limitations, models enable complex phenomena to be simulated. Most soil water flow problems cannot be solved or evaluated by any other means, so a computer model which can be operated rapidly using easily obtainable data is an invaluable tool for soil scientists (Hutson, 1983).

4. SOIL MOISTURE AND SOLUTE MOVEMENT IN THE ROOT ZONE

To gain an understanding of the hydrosalinity processes within the root zone and to gain information on the quantity and quality of water draining into the delivery zone, two orchard fields under micro-jet irrigation were monitored. At three sites within each of these orchards the matrix potential and soil moisture chemistry were monitored on a daily and weekly basis respectively. These data are also used in Chapter 5 for an assessment of several soil moisture and solute transport models.

4.1 THE MICRO PLOTS

4.1.1 Selection of the micro-plot study sites

Initially it was decided to locate the micro-plots at sites representative of the major irrigation techniques found within the lower Sundays River irrigation area and if possible on areas representative of the major soil types. To help select these sites a geomorphic map of the Coerney valley was produced (figure 2.5). This map indicates a close relationship between the surface soil units of the lower Coerney valley (figure 2.6) and its alluvial terraces.

However, an examination of the alluvial profiles, through which boreholes have been drilled, shows that the alluvium comprising each terrace to be highly variable, both spatially and with depth. This prevented the demarcation of areas with homogeneous soil characteristics which would have enabled the plot studies to be located at sites representing definable areas within the Coerney valley. Therefore as the irrigated soils of the Coerney valley are predominantly of the Oakleaf form, it was decided that the micro-plots should be located on this soil type.

The degree of assistance that could be gained from the local farming community became the major factor in deciding the location of the plot studies. It was therefore decided to establish two sites representative of areas under micro-jet irrigation on the farm Daisy Dell where the project was assured of good cooperation from the farmer and some assistance with data collection. This has proved essential for the collection of daily soil moisture data owing to the distance between the field area and Rhodes University. Orchards D and M of this citrus farm were selected as they are located on different terraces and although they have soils of the same form, they have different textural characteristics (figure 4.2 and table 4.1).

4.1.2 Orchards D and M of Daisy Dell farm

Orchard D is located on a low terrace with a dark grey silty clay loam. Orchard M is located on a higher terrace with a soil that is predominantly a reddish brown sandy clay loam, over a loamy sand. Orchard D carries a stand of lemon trees planted in 1977, whereas orchard M has a stand of Delta orange trees planted 1974 while those in orchard D have a spacing of 2.75 metres in rows 6.5 metres apart. The trees in orchard M are at a spacing of 3 metres and in rows 6.5 metres apart. Two micro-jet sprinklers are located 0.5 metres on either side of the trees and irrigate an arc away from the tree itself. Each pair of micro-jets irrigate an area of approximately 12 sq metres at a rate of 5 mm per hour. The rate of irrigation was regularly checked and the system well maintained by the farmer. It is therefore reasonable to assume that the rates of irrigation across both orchards are fairly uniform and comparable. The farmer has maintained a reliable record of the irrigation and fertilizer applications for these orchards.

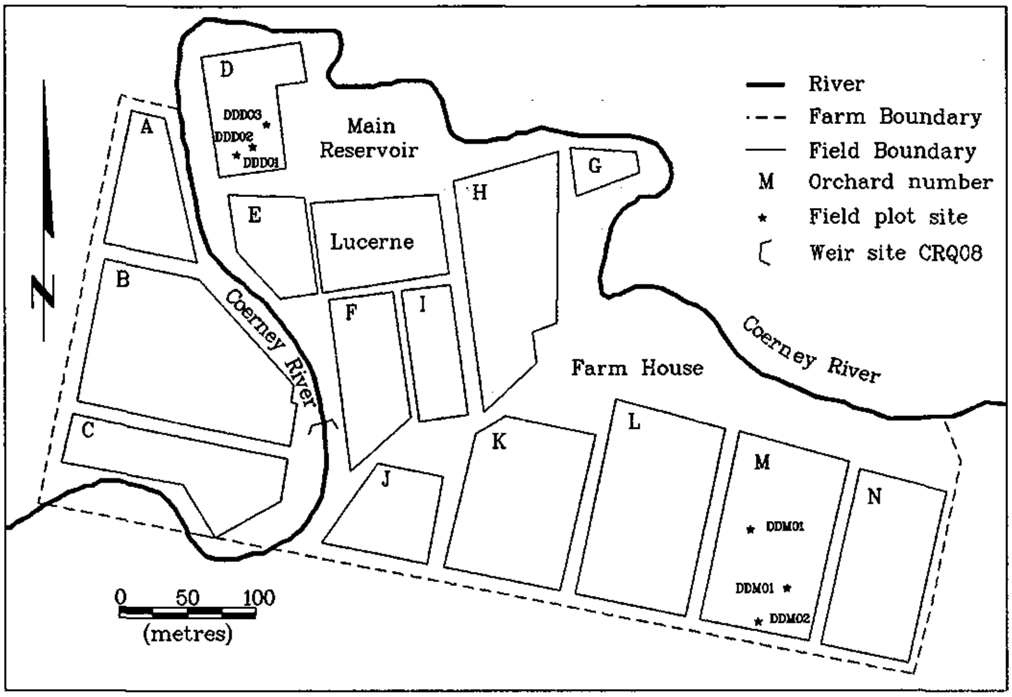


Figure 4.1 Lay out of orchards on Daisy Dell farm.

4.1.3 Selection of study sites within orchards D and M

The micro-plot studies were established to provide representative information on the movement of soil water and solutes under irrigation and on the quantity and quality of soil moisture draining into the delivery zone. It was therefore important to gain some information on the spatial variability of soil moisture within each orchard to ensure that representative sites were selected for detailed instrumentation. This was done by carrying out a soil moisture survey within each orchard.

The soil profile was sampled at three depths, 30, 60 and 110 cm, representing the top profile, the base of the root zone and beneath the root zone respectively. Soil moisture data were obtained using the gravimetric method. The sampling density of the soil profile was limited by the time available, the necessity to limit disturbance to the soil and tree roots, and by the area of each orchard. Orchard D, with 10 rows of a maximum of 36 trees, was sampled at intervals of 10 trees to give a total of 31 profiles. Orchard M, with 13 rows of 54 trees, was sampled at a spacing of 13 trees to give a total of 48 sample profiles. The sitting of each profile was standardised so as to eliminate moisture variations due to distance from the tree base and micro-jets. Each profile was located midway between trees at the edge of the canopy. Working beneath the low tree canopy would have been difficult and led to fewer profiles being sampled in the available time.

The soil moisture data at the 30 cm depth, combined 30 and 60 cm depth and combined 30, 60 and 110 cm depths were ranked in ascending order. Those values close to, or bracketing the mean were then determined as were those with soil moisture values one standard deviation either side of the mean. Three plot sites within each orchard were then selected with one at the mean soil moisture and one either side of the mean. It was hoped that the selected sites would provided representative soil moisture information for the two respective orchards. However, it should be noted that the site selection was based on sample means and standard deviations. In orchard M where the soil moisture data approximated a normal distribution the selected sites should be representative of the orchards soil moisture regime. Whether this remains true for orchard D, for which the moisture data are strongly skewed is open to question. Within orchard D sites DDD02 and DDD03 were selected as representative of soil moisture conditions one standard deviation drier and wetter than the mean respectively (figure 4.1). Site DDD01 is representative of mean soil moisture conditions. In orchard M site DDM03 is representative of the mean conditions and sites DDM01 and DDM02 represent dry and wet conditions respectively.

4.1.4 Instrumentation of the micro-plots

Each micro-plot was instrumented with a bank of tensiometers installed at depths of 30, 60, 90 and 120 cm as well as in situ soil moisture extraction tubes installed at depths of 15, 30, 60, 90 and 120 cm. Neutron access tubes were installed at sites DDD02 and DDD03 in orchard D and sites DDM01 and DDM02 in orchard M to enable the tensiometers to be calibrated and for monitoring soil moisture below the root zone. At site DDM03 in orchard M an array of neutron access tubes was installed to examine the spatial variation of soil moisture around an individual tree. In orchard D a gravel layer at a depth of 2 metres limited the depth to which the access tubes could be installed. In orchard M the tubes were inserted to a depth of 3 metres. Volumetric soil samples were collected as the holes for the neutron access tubes were being made. These soil samples were analysed to determine their bulk density and textural characteristics (table 4.1).

The tensiometers were read on a daily basis from June 1989 until August 1990. From that date until June 1991 the tensiometers were read three times per week. Soil moisture samples were collected on a weekly basis and analysed for their major anions and cations, electrical conductivity and alkalinity. Soil moisture at each micro-plot was determined at 30 cm depth intervals on a weekly basis using a neutron probe. Rainfall data were recorded for a site on Daisy Dell farm in close proximity to the two study orchards and A - pan evaporation data were recorded by the Dept. of Agriculture at their Addo Citrus Research Station.

Maximum evaporation rates occurred during the summer months from November to February (figure 4.2). However, most irrigation was applied between January and April. Irrigation and evaporation maxima do not coincide as most irrigation is required during the period of fruit growth and ripening. Minimum evaporation rates occur during the winter months from April to August, when there is correspondingly less irrigation.

Table 4.1 Textural characteristics of soils at sites DDD01 and DDM03.

Depth (cm)	SITE DDD01 % sand	SITE DDD01 % silt	SITE DDD01 % clay	SITE DDM03 % sand	SITE DDM03 % silt	SITE DDM03 % clay
0 - 30	50	19	31	66	6	25
30 - 60	51	17	32	60	13	27
60 - 90	68	12	20	65	12	23
90 - 120	66	12	22	74	9	17
120 - 150	71	10	19	80	7	13
150 - 180	85	5	10	79	7	14

Other than a drop in salinity immediately following the flood event of November 1989, electrical conductivity of the irrigation water supplied to lower Coerney valley has remained relatively stable. However, figure 4.3 does illustrate an increase in the concentraion of sodium, magnesium and chloride during the 1989-90 irrigation season.

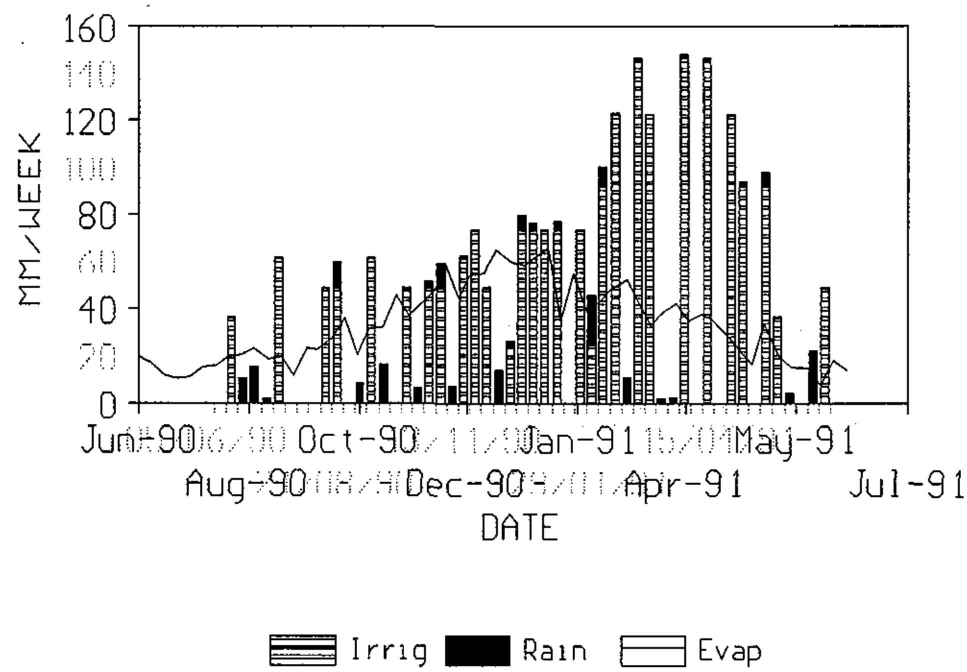


Figure 4.2 Weekly irrigation, rainfall and evaporation used for the micro-plot study.

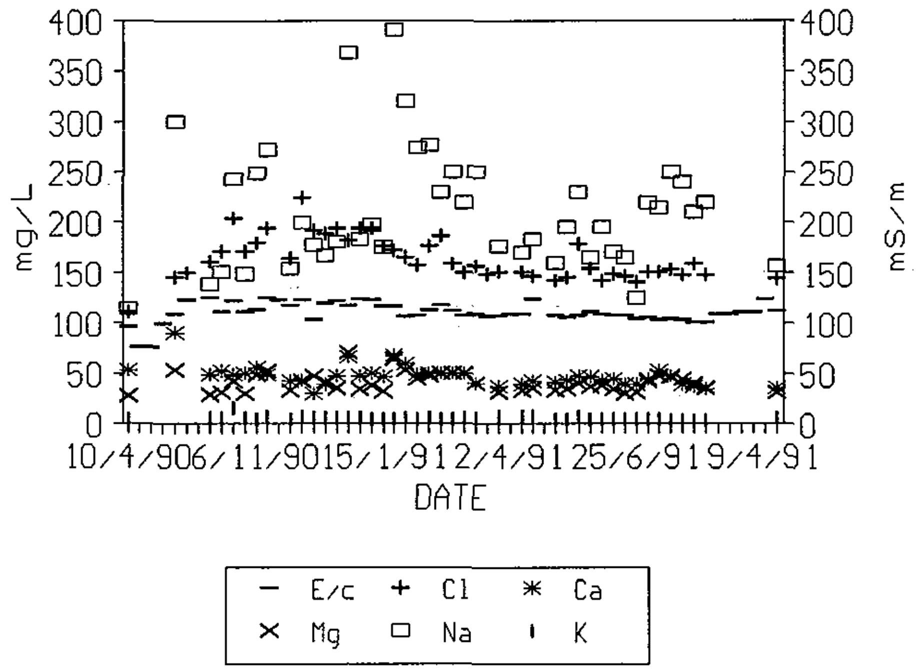


Figure 4.3 Irrigation water quality for the period 10-4-90 to 10-4-91.

4.2 SOIL MOISTURE

Weekly tensionmeters measurements for DDM03 were converted to measures of soil moisture using a relationship derived for this site between matric potential and gravimetric soil moisture values. Figure 4.6 shows that on average the 30cm layer is more moist than the deeper layers and exhibits the greatest variability, which is clearly indicated by its large standard deviation of 1.49 (table 4.2). This variability could be expected as this layer forms the soil surface, and is therefore more affected by the boundary effects of evaporation, irrigation and rainfall than the underlying layers. With increasing depth the soil moisture decreases and becomes more uniform through time.

4.3 DETERMINATION OF SOIL MOISTURE FLUX

Three different approaches are adopted for the measurement of soil moisture flux. Initially weekly values were determined using a Darcian approach. However, as these values seemed unrealistically small, soil moisture flux was also determined using the water budget and a chloride mass balance approach.

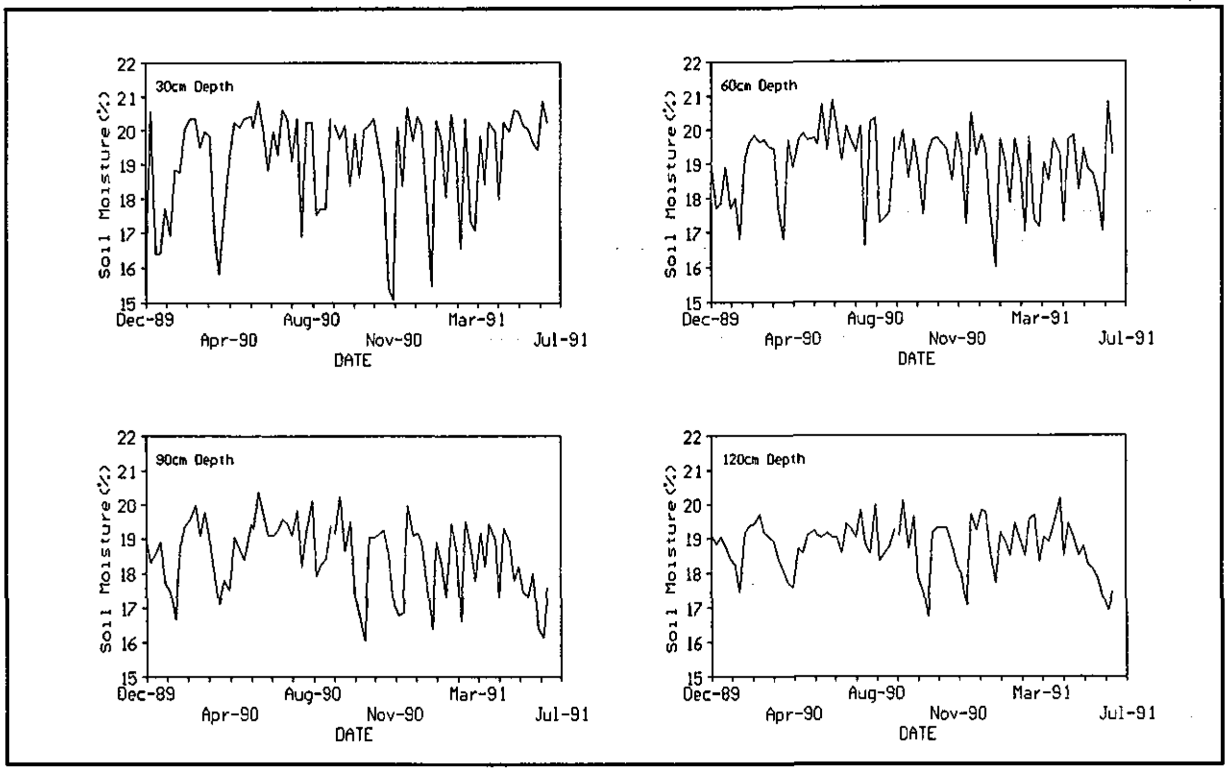


Figure 4.4. Soil moisture variations at micro-plot DDM03 on Daisy Dell farm.

Table 4.2. Summary of soil moisture percentages by volume at micro-plot DDM03 for the study period.

	30 cm	60 cm	90 cm	120 cm
Mean	19.14	18.94	18.51	18.79
Minimum	15.08	16.00	16.07	16.73
Maximum	20.89	20.89	20.36	20.16
Std. Dev.	1.49	1.10	1.05	0.73

4.3.1 Determination of soil moisture flux using Darcy's law

The Darcian approach for estimating soil moisture flux involves determining the product of the soils hydraulic conductivity and the hydraulic gradient (Eq. 4.1). The hydraulic gradient is the difference in matric potential for the time over which the soil moisture flux is to be determined.

$$q = -K.gradH \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (4.1)$$

where

q	=	moisture flux
K	=	soil hydraulic conductivity
gradH	=	hydraulic gradient

(Vachaud and Vauclin, 1990)

The flux values determined using this technique are extremely small with a maximum of only 30mm/week at the 60 cm depth (figure 4.7). This is most unrealistic when one considers that for some weeks during the irrigation season up to 150mm of water is applied at DDM03. Figure 4.5 also shows the soil moisture flux at the 60 and 90cm depths to be much greater and more variable than the 30 and 120cm layers. This finding is also unrealistic when one considers that the irrigation water is applied to the surface and that water in the 60 and 90 cm layers must pass through the 30 cm layer.

4.3.2 Determination of Soil Moisture Flux using the Water Budget Approach

As a consequence of the unrealistic values of soil moisture flux determined using the Darcian approach, flux values were also determined using a water balance method. This approach determines the depth of water that passes through a given layer of soil for a nominated time period by compiling the following water balance:

$$q = I + R - Et - dS \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (4.2)$$

where	q	=	moisture flux (mm/day)
	I	=	depth of irrigation (mm)
	Et	=	depth of evapotranspiration (Pan Evaporation x Crop Factor of 0.7)
	dS	=	change in soil moisture storage (mm)

The flux rates determined by this method are illustrated in figure 4.8. The irrigation season from January to April is clearly reflected at all depths by the increased rates of soil moisture flux. However, the estimation of similar flux rates at all depths through the soil profile is unrealistic. One would expect flux rates to decrease with depth as the amount of moisture within the soil profile decreases with root uptake. Rates of soil moisture flux should also decrease with depth as the hydraulic conductivity of the soil decreases through compaction and increased bulk density. The flux rates estimated using the water balance approach are also considered suspect as figure 4.6 shows considerable upward movement (negative fluxes) of moisture following each irrigation application. The estimated upward fluxes often exceed 50mm per week, which is totally unrealistic especially as the water table at approximately 7 metres below the surface is too deep to provide a source of moisture for upward capillary movement.

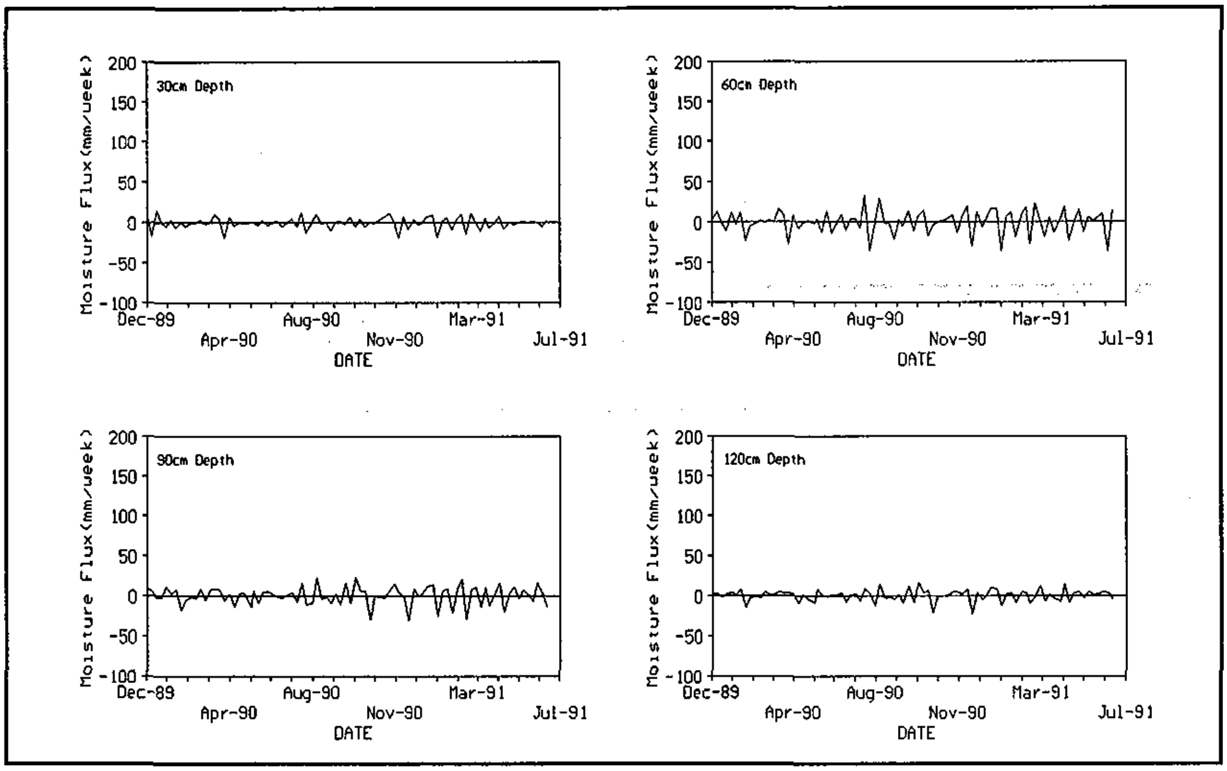


Figure 4.5 Soil moisture fluxes at micro-plot DDM03 determined using the Darcian approach (-ve fluxes represent upward movement).

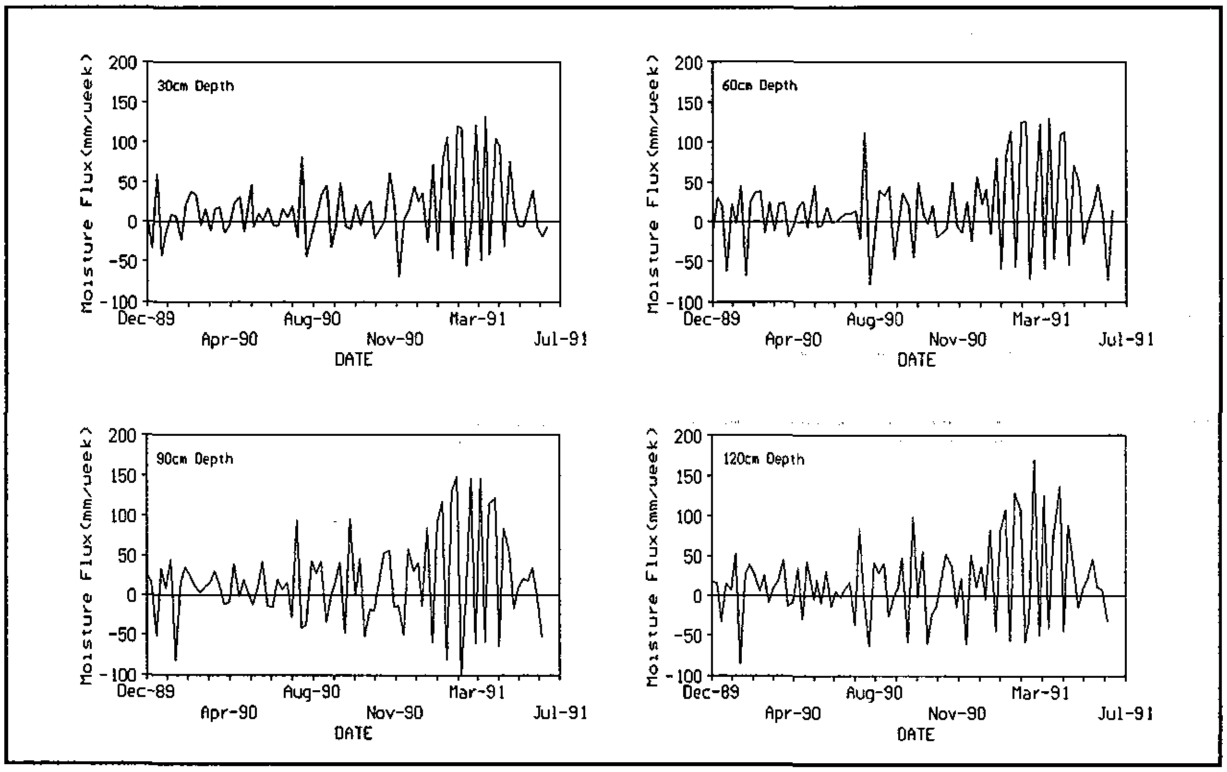


Figure 4.6 Soil moisture fluxes at micro-plot DDM03 determined using the water budget approach (-ve fluxes represent upward movement).

4.3.3 Determination of soil moisture flux using a Chloride mass balance

A volumetric mass balance approach was applied to an area of $10cm^2$ in which the soil moisture occupying each 30cm depth was translated into a volume of water within the defined $3000cm^3$ block of soil. The solute load for each ion in this moisture was calculated from the soil solution concentrations measured at weekly intervals.

The net solute loss from each 30cm layer was then determined for each ion and compared with the salt load from the incoming irrigation water (table 4.3). The close approximation between the load of each ion within the resepective layers and the net incoming load of that ion in the irrigation water suggests that this technique is accounting for most of the moisture movement within the profile. The residual between the irrigation input load and the load calculated using the mass balance was calculated from:

$$\text{Residual} = \frac{\text{(irrigation input - load in the layer most different to the irrigation input)}}{\text{(Irrigation input)}} * 100$$

The similarity between the solute load input through irrigation and the solute loads lost from each 30cm soil layer is shown by the residuals in table 4.3. Potassium has the greatest residual value between incoming irrigation load and outgoing load in the soil solution. This may be due to plant uptake, as K^+ is a constituent of fertiliser and hence a plant requirement. Cation exchange processes will also affect K^+, leading to its preferential adsorption over hydrogen ions. It is likely however, that the extremely low concentrations of this ion result in a proportionately greater error margin in field collection and laboratory analysis. Calcium shows the second greatest disparity between incoming load and calculated outgoing load. This may also be due to cation exchange processes, especially as Ca^{2+} is added to the soil as an ameliorant for dispersed soils.

Chloride shows an unexpectedly high residual of 8.42%. As a conservative ion, Cl^- is less likely to become involved in adsorption onto soil exchange sites or react with other ions present in the soil solution. For this reason, Cl^- is commonly used as a tracer for soil moisture movement in hydrosalinity modelling. The residual indicates approximately 8.42%

Table 4.3 Solute load of irrigation water and that lost from each $3000cm^2$ soil layer down the profile at microplot DDM03 for the study period.

	30 cm	60 cm	90 cm	120 cm	Irrig. Water	Residual (%)
Chloride (mg/l)	4660	4544	4285	4285	4680	8.42
Calcium (mg/l)	1572	1503	1453	1414	1622	12.85
Magnesium (mg/l)	1174	1160	1144	1127	1184	4.8
Sodium (mg/l)	5899	5825	5716	5995	5995	4.64
Potassium (mg/l)	136	111	98	93	149	37.43

of the load coming into the soil through the irrigation water is not accounted for by the load calculated by the mass balance technique. This percentage is low relative to the uncertainty in measurement, and variability problems in the unsaturated zone. However, two cations show smaller residuals, and thus a closer correlation between irrigation and calculated loads.

Magnesium shows an unexpectedly low residual of 4.8%. Divalent cations are susceptible to cation exchange processes and reactions within the soil solution. Mg^{2+} is therefore not considered to move freely through the soil in response to gravitational and capillary forces alone. The low residual indicates that this not is occurring, which can be explained if the soil is saturated with Mg^{2+}. As such a situation is unlikely a further possible explanation is that errors in measurement were able to counteract each other, and result in a unusually low residual.

Sodium is shown to have the lowest residual between irrigation input and calculated loads, 4.64%. The high concentrations of this ion in the soil has been cited as a causal factors for the poor drainage experienced in some parts of the lower Coerney irrigation area (Folscher, 1981). Remedial measures including the application of Calcium in the form of gypsum ($Ca.SO_4.2H_2O$) to displace the Na^+ ions from the cation exchange sites and encourage flocculation have been routinely applied throughout the entire Sundays River valley. It is feasible therefore that the soil system may be saturated with Na^+ ions, such that these ions can move as freely through the soil as a non-reactive species.

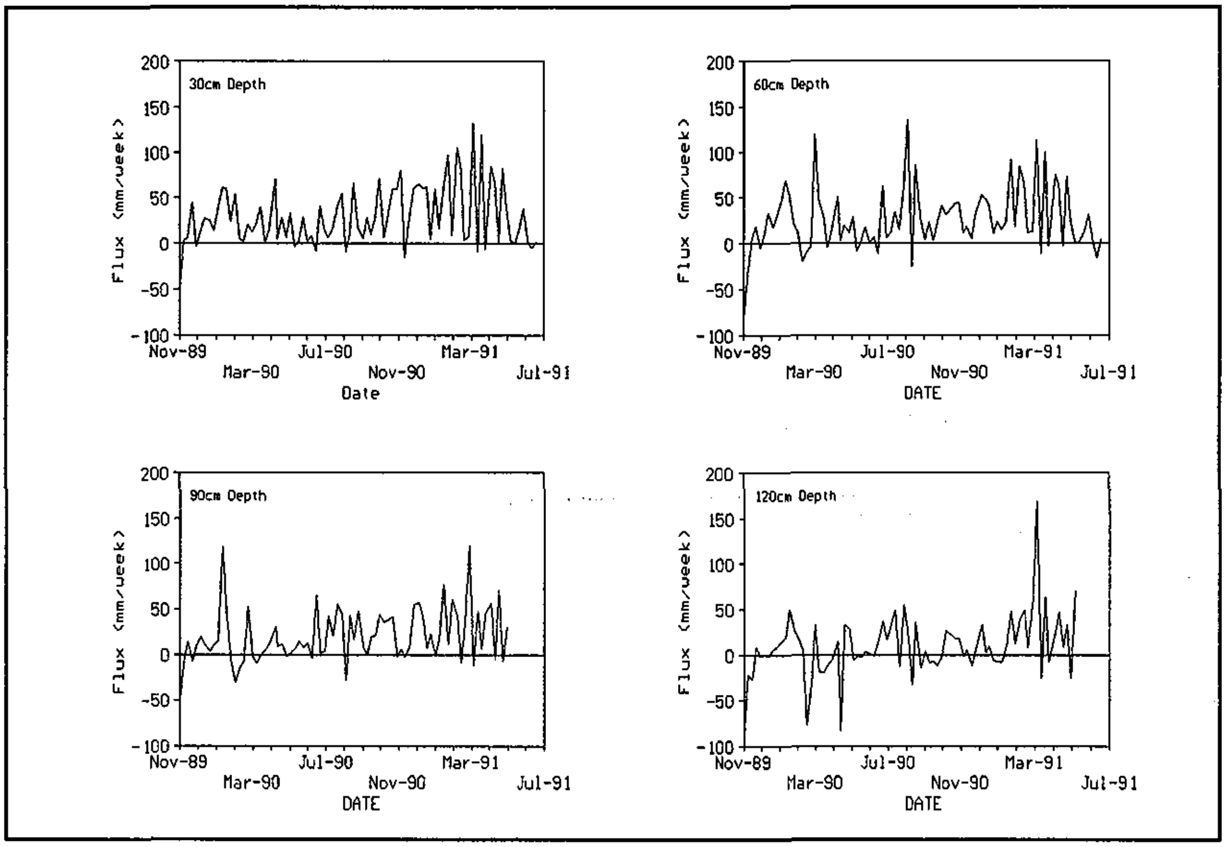


Figure 4.7 Soil moisture fluxes at micro-plot DDM03 determined using the Chloride mass balance approach.

Although lower residuals were obtained using Magnesium and Sodium loads, Chloride was considered to be the most likely to portray soil moisture fluxes without interference effects due to cation exchange and solute interactions. Weekly values of soil water Chloride concentrations (mg/L) were divided by the load calculated each week (mg), to provide an estimate of moisture flux in litres. This amount was then converted to millimeters, to allow comparison with the other methods of flux calculation.The fluxes illustrated by figure 4.7 appear to be more realistic than shown in figures 4.5 and 4.6. The mass balance generated fluxes are of an acceptable order of magnitude, considering the inputs from irrigation and rainfall shown in figure 4.4, unlike the fluxes generated using differences in matric potentials (figure 4.5). There are less extreme negative fluxes in figure 4.7, and a discernible reduction in flux with depth from 30 to 120cm, unlike the flux determined using the water balance technique (figure 4.6).

4.3.4 Evaluation of methods of soil moisture flux determination

A summary of the soil moisture fluxes generated using each technique is presented in table 4.4. Any short term erroneous fluctuations are less influential when an average rate for the entire period is considered. The method of flux determination using differences in matric potential shows upward movement above 60 cm, below which downward movement predominates. As the upper layers provide the moisture which moves into the lower layers, this data seems erroneous.

Both the water balance and Chloride mass balance approaches produce more realistic results than the Darcian flow method. Overall flux determined using these two techniques decreases with depth, as is expected due to the losses of evapotranspiration, and potential lateral flow within the soil matrix. All fluxes are positive, which is also to be expected under conditions of frequent irrigation (figure 4.2).

The confidence in the mass balance method of flux determination led to a statistical comparison with the fluxes generated by the other techniques. Variation in flux determined

Table 4.4. Mean weekly soil moisture flux (l/week) at micro-plot DDM03.

Method of Est.	Depth				R^2 against mass balance method
	30 cm	60 cm	90 cm	120 cm	
Darcian	-7.7	-6.2	12.81	4.55	0.54
Water Balance	1.17	1.11	1.07	1.01	0.59
Mass Balance	2.47	2.09	1.42	0.70	1.00

where :

Darcian	=	flux estimated using Darcy's Law
Water Balance	=	flux estimated using a soil water balance
Mass Balance	=	flux estimated using a chloride mass balance

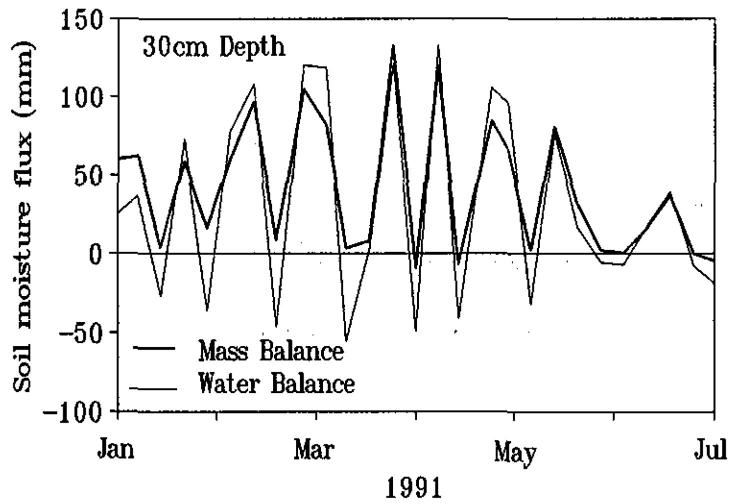


Figure 4.8 Comparison of soil moisture flux determined using the water balance and Chloride mass balance techniques at the 30 cm depth for the period January to July 1991.

using the water balance only accounts for 54% of the variation in the flux determined using the mass balance method, while the water balance accounts for 59% of the variation in flux generated using the mass balance. This means that the Darcian flow method is unable to replicate more than 54% of the variability, as well as the absolute amount of flux generated by the mass balance approach. The water balance method may have more realistic overall fluxes, but the variability in flux determined using the water balance only replicates 59% of the variability in flux determined using the mass balance approach.

The failure of the matric potential technique to account for more than 54% (table 4.6) of the soil moisture movement determined using the mass balance may be explained in terms of the physical processes operating within the soil and the different theoretical bases underlying each technique. Differences in matric potential do not account for absolute moisture movement, as rapid throughflow of the mobile phase may occur too quickly for the tensiometer to register the correct soil potential.

The moisture held in the soil by capillary forces that results in the tension measured is largely micropore water. Water in macro-pores is governed by gravitational forces, which result in rapid draining of water. As the tensiometer adjusts to ambient soil moisture over a period of several hours, in that time a considerable amount of soil moisture could have moved through the macropores within the soil. The unsaturated conductivity of the soil ranges from 5-12 mm/day movement within the soil matrix is slow enough to be gauged by the tensiometer. On the basis of these results it is suggested that the matric potential technique measures micropore flow.

Lateral flow is not accounted for by any of the three techniques considered. The mass balance method considers a 10cm^2 area which technically allows for a small amount of variability in the lateral direction, but the soil moisture measurement taken by the tensiometer is considered constant throughout the 3000cm³ soil volume. Lateral variations in soil moisture or concentration are not taken into account. The process of lateral flow is to be expected, although no evidence has been collected to support the occurrence of this phenomenon. Lateral flow would account for the systematic underestimation of load lost by the chloride mass balance technique against irrigation input with depth for all ions. With depth, soil physical properties such as bulk density, compaction and permeability will change to inhibit vertical movement in favour of lateral flow.

Under a microjet irrigation situation in the orchard, there will be wetted areas immediately under irrigation emitters and dry areas between the tree rows. The matric potential in the dry soil will be higher than the wet soil, so moisture will be drawn towards the dry areas by the hydraulic gradient set up in this way. Thus the irrigation technique further encourages lateral flow within the soil.

4.4 SOIL WATER CHEMISTRY

Soil solute concentrations were measured at each depth (figures 4.9 to 4.11), and are summarised in table 4.5. Chloride (figure 4.9) and Sodium (figure 4.11) vary sympathetically over time, and exhibit similar variability at all depths. The concentration of these ions increases with depth to 120cm. This suggests the soil peak of Sodium and Chloride lies below the root zone, and beyond the depth investigated.

Chloride is a conservative ion, unreactive and unlikely to take part in exchange processes. The Cl^- ion concentration in the irrigation water shown in figure 4.5 is nearly constant. Given that the amount of irrigation varies over time, the concentrations of Chloride in the soil are

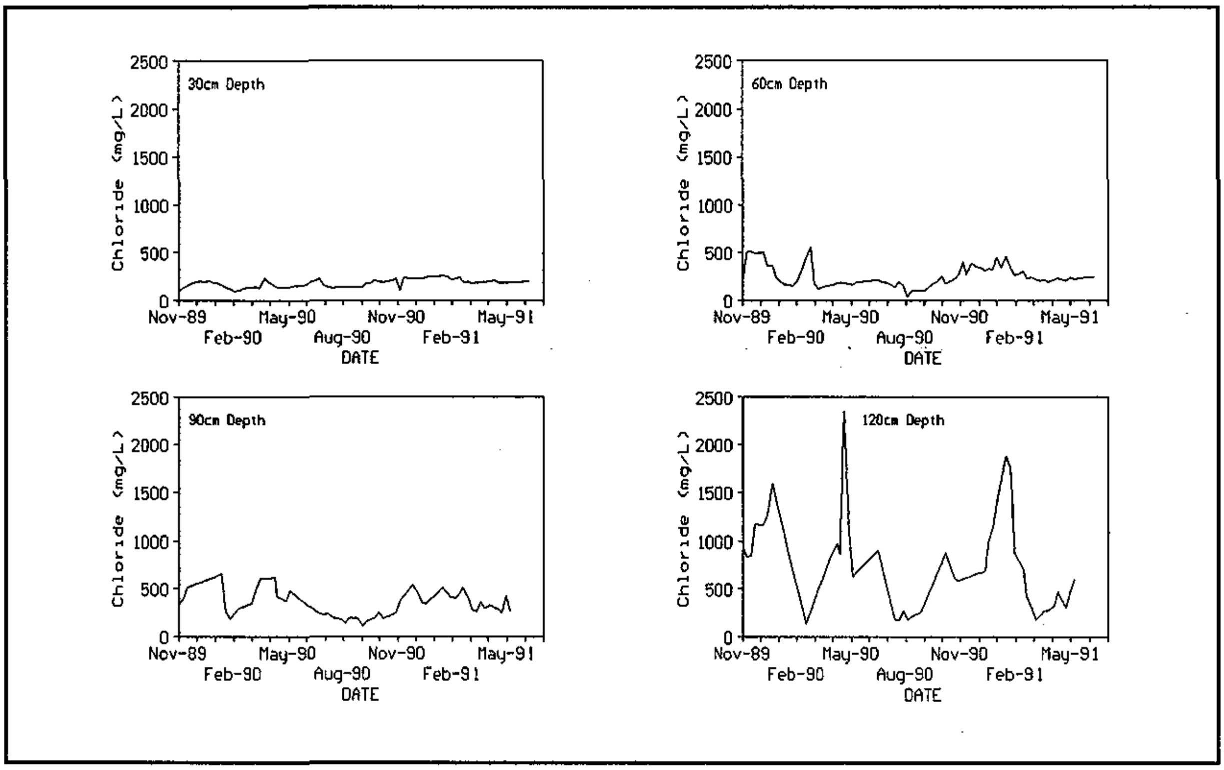


Figure 4.9 Weekly soil water Chloride concentrations at micro-plot DDM03.

Table 4.5 Summary of Soil Solutes measured at Daisy Dell micro-plot over the study period

Solute	Maximum (mg/l)	Minimum (mg/l)	Mean (mg/l)	Standard Deviation
Chloride				
30cm	259	89	175	42
60cm	558	35	247	113
90cm	657	106	367	142
120cm	2351	139	723	433
Calcium				
30cm	174	51	104	29
60cm	380	69	147	66
90cm	136	42	88	21
120cm	629	36	152	90
Magnesium				
30cm	102	38	58	15
60cm	213	47	89	36
90cm	100	29	60	16
120cm	355	28	97	48
Sodium				
30cm	400	115	185	49
60cm	640	151	278	107
90cm	914	262	553	146
120cm	2160	250	1094	508
Potassium				
30cm	205	0	35	21
60cm	113	4	57	21
90cm	30	6	21	4
120cm	49	3	14	9

expected to show less variability than the other ions, which are susceptible to cation exchange processes, and also show variability in the irrigation water before entering the soil. However, table 4.5 shows that Cl^- has the greatest standard deviation at all depths, except for Sodium. It is suggested that the Chloride load in the irrigation water is not the only source of Cl^- in the system. As the soil has developed on alluvial material, derived from marine sediments,

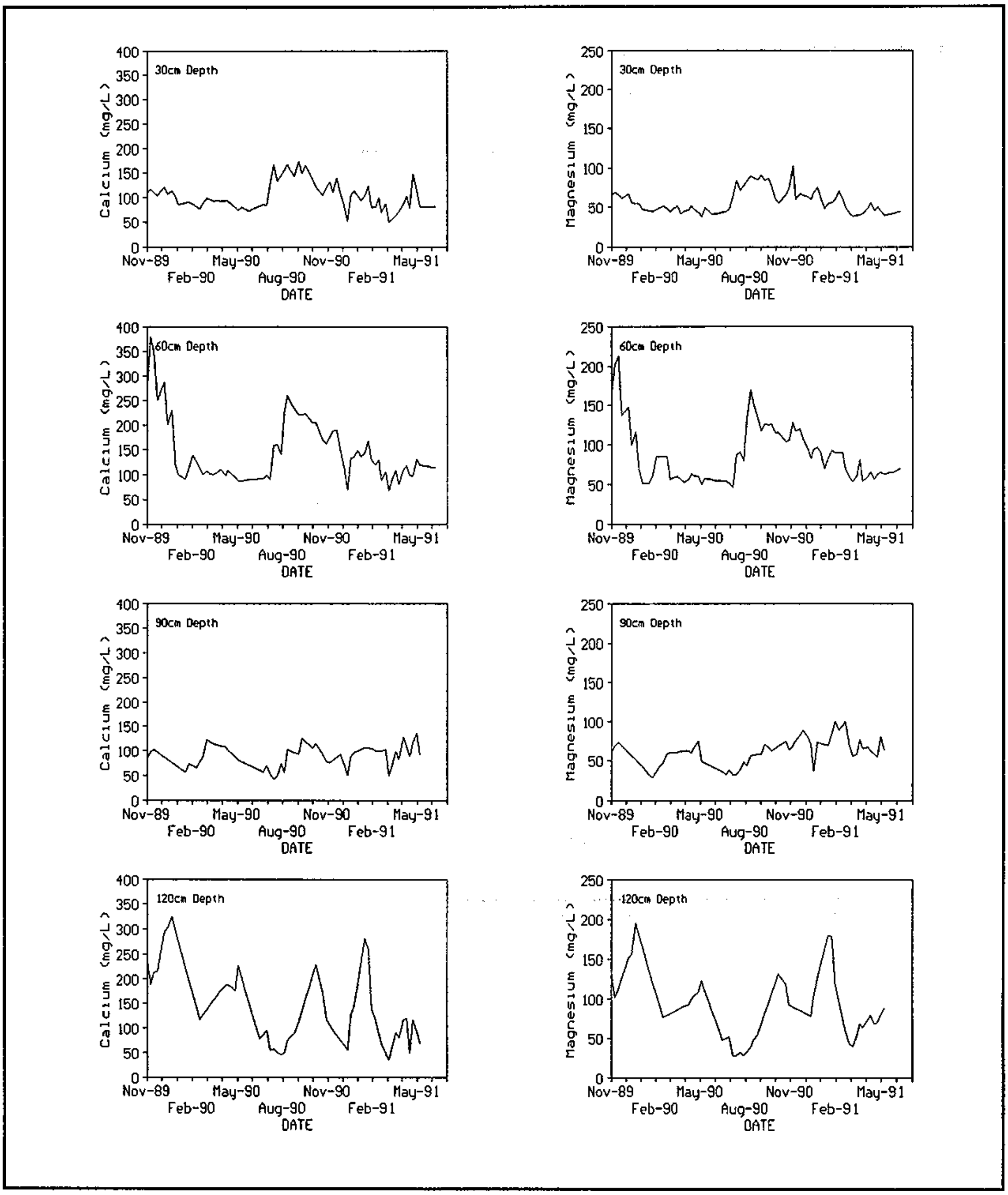


Figure 4.10 Weekly soil water Calcium and Magnesium concentrations at micro-plot DDM03.

the soil minerals are likely to contain high concentrations of both Cl^- and Na^+, which may become mobilised during leaching processes. Calcium and Magnesium are closely related in both variability and absolute concentration (figure 4.10). The concentration peaks of these cations occur at the 60cm depth, below which there is a gradual decline in concentration. The Potassium cation shows a lower absolute variability than all other ions and also exhibits

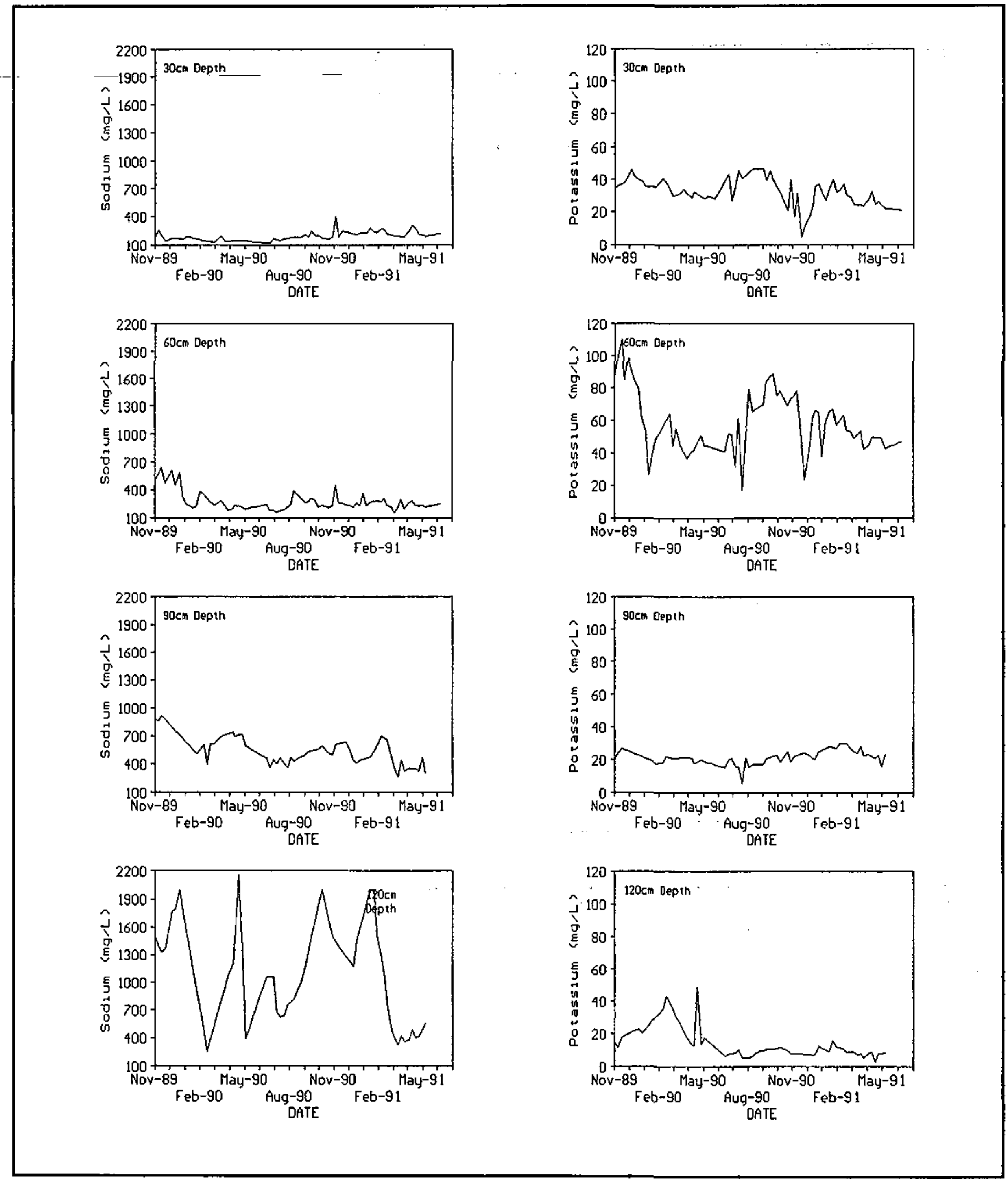


Figure 4.11 Weekly soil water Sodium and Potassium concentrations at micro-plot DDM03.

a concentration peak at 60cm, with a marked decrease in concentration below this level (figure 4.11). The Orchard rooting depth was observed to average 60cm, which may be significant in influencing the concentration profiles shown in figure 4.9 to 4.11, especially Ca^{2+}, Mg^{2+} and K^{+}. Selective uptake of soil moisture and nutrients by plant roots serves to concentrate the remaining soil solution, and this may be more significant where concentrations are low, as with Potassium.

4.5. SOLUTE FLUX

The convective loss of ionic species was determined from:

$$Li = Ci.q.t \qquad (4.3)$$

where Li = leaching of ionic species i
Ci = concentration of species in the soil solution
q = moisture flux (mm/week according to the mass balance technique)

Solute fluxes of Chloride, Calcium, Magnesium, Sodium and Potassium are illustrated in figure 4.12 to 4.16. The Cl^{-} fluxes are consistently greater than those for the analysed cations. This is due to its overall higher absolute concentration (shown in figure 4.9), higher solubility and hence mobility. Chloride is the least reactive ion considered, and so less likely to be taken up in exchange reactions within the soil. Sodium flux is the next highest, which is also attributed to its high natural concentration.

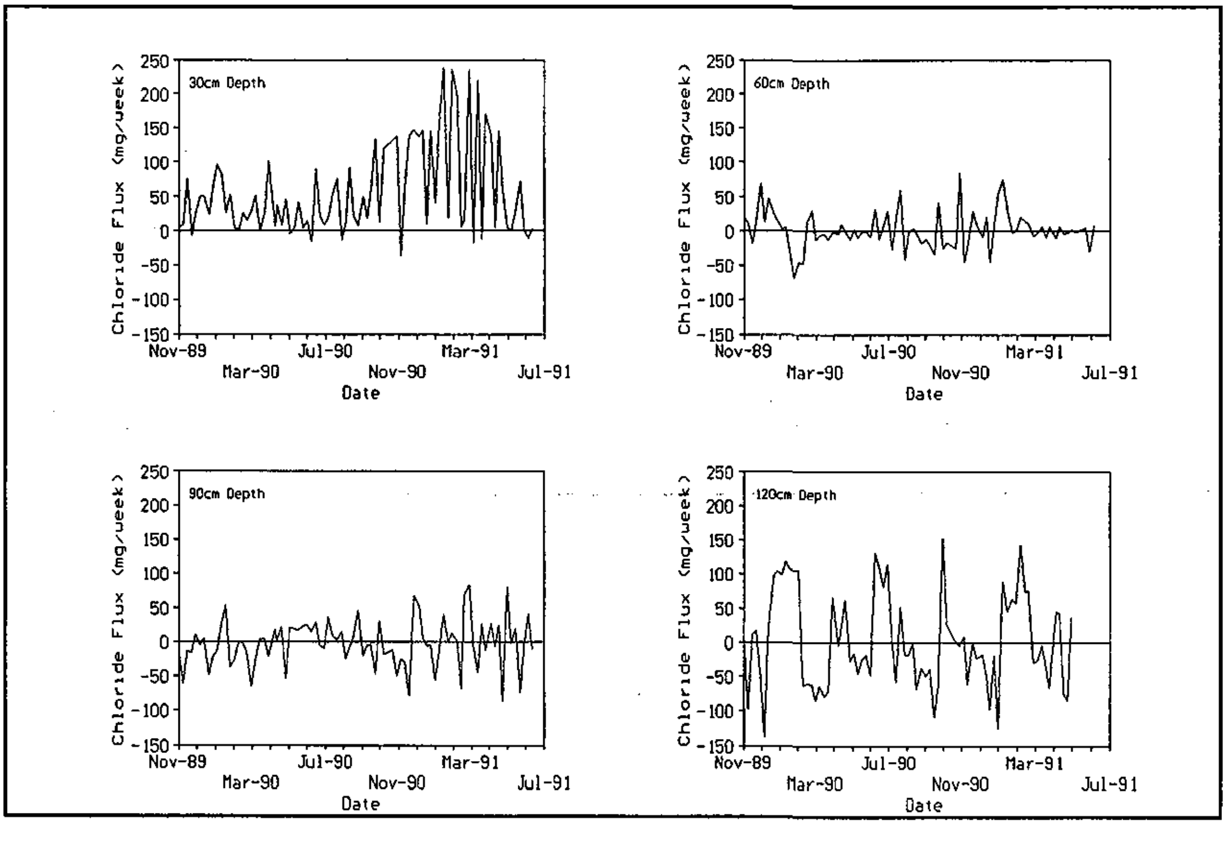


Figure 4.12 Weekly Chloride fluxes determined for micro-plot DDM03.

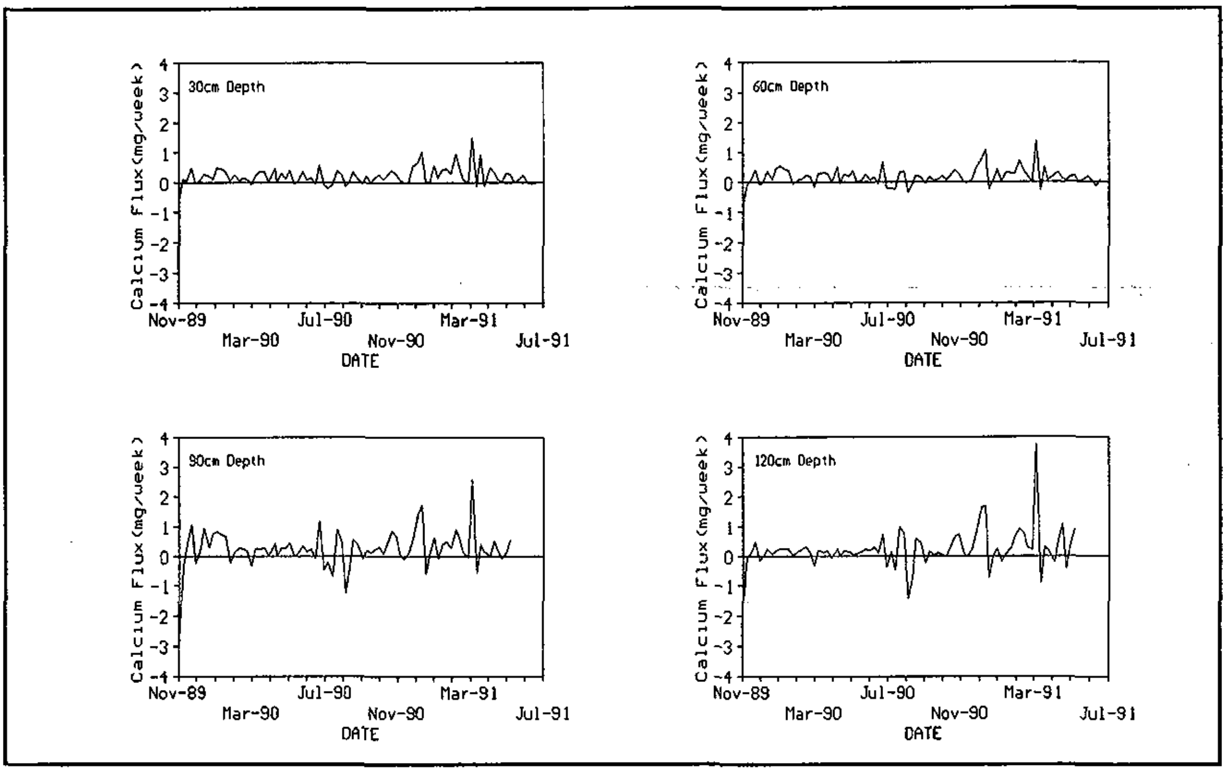


Figure 4.13 Weekly Calcium fluxes determined for micro-plot DDM03.

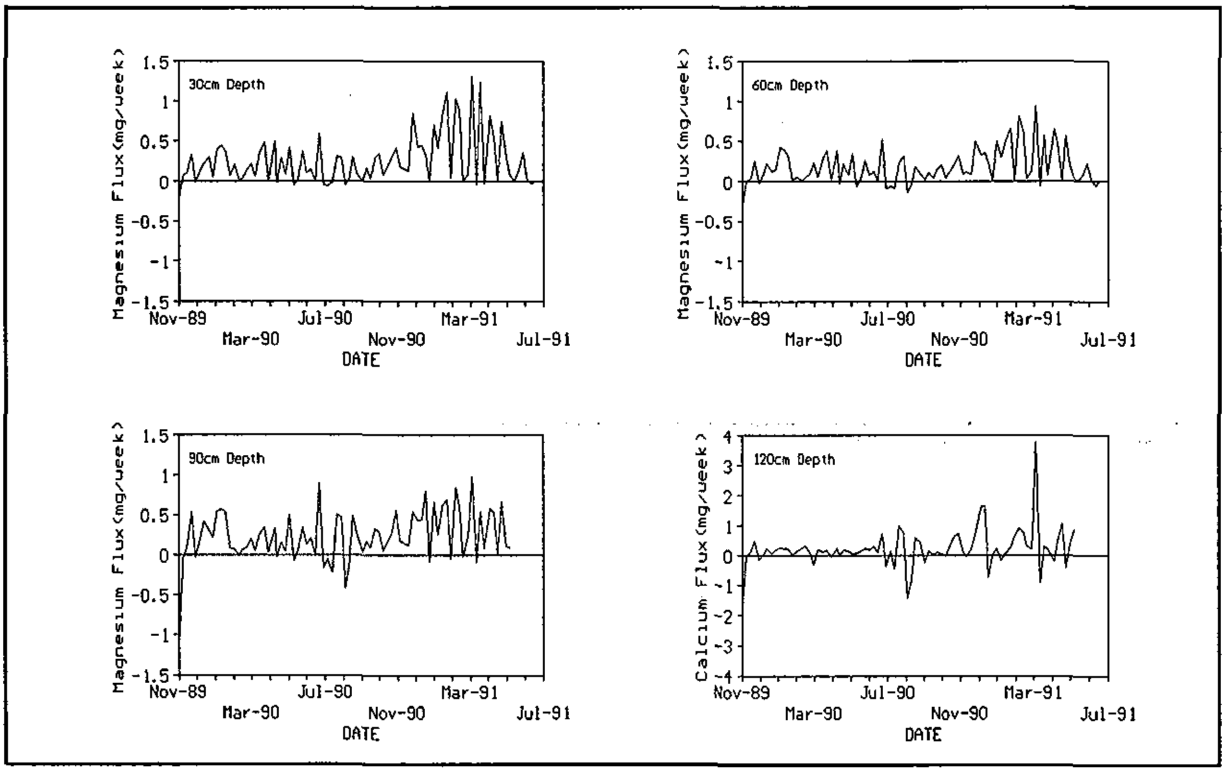


Figure 4.14 Weekly Magnesium fluxes determined for at micro-plot DDM03.

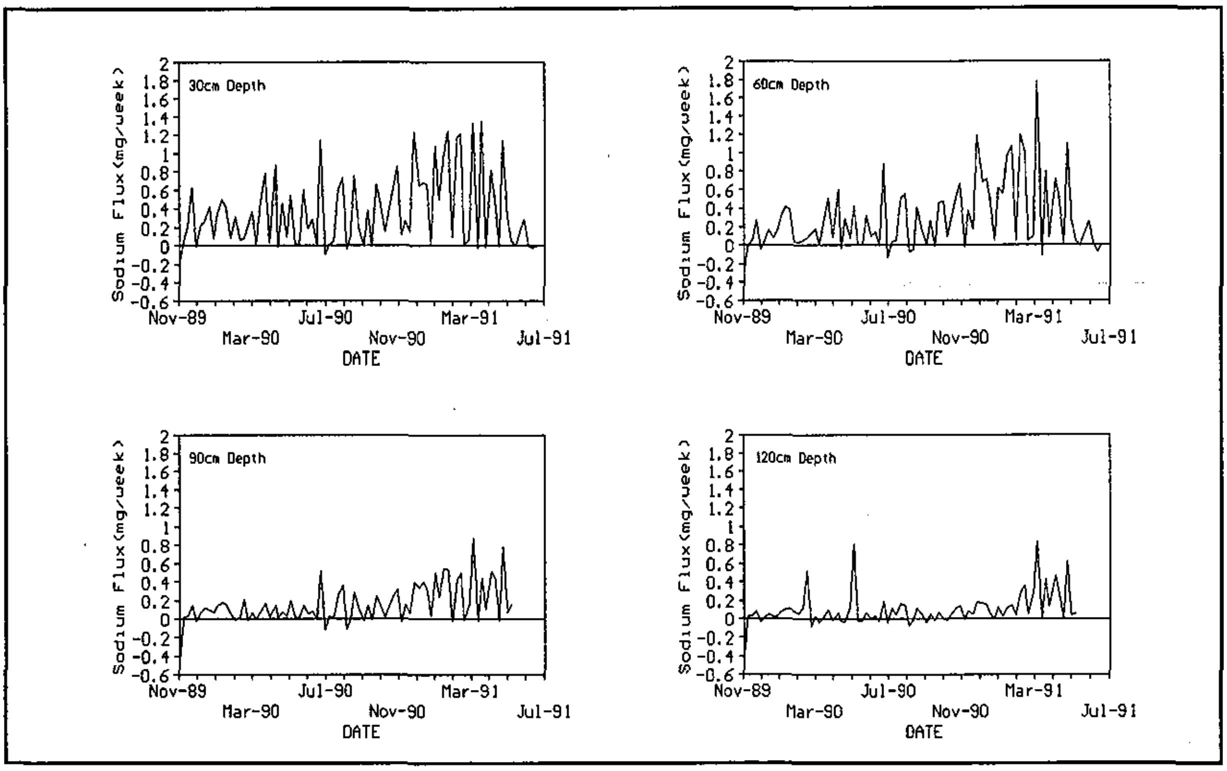


Figure 4.15 Weekly Sodium fluxes determined for micro-plot DDM03.

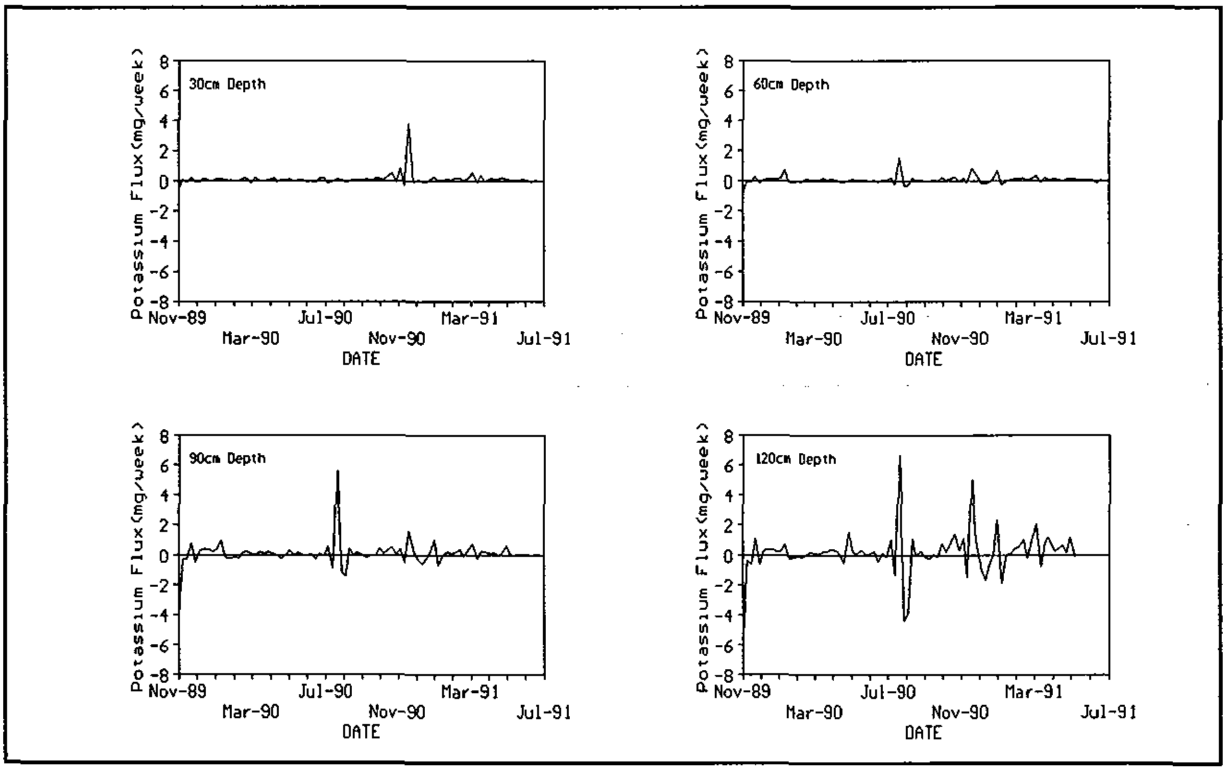


Figure 4.16 Weekly Potassium fluxes determined for micro-plot DDM03.

The high Calcium concentration may be explained by the history of gypsum ($Ca.SO_4.2H_2O$) application in the orchard, although none was added during the study period. Previous applications were at a rate of 4 and 2 tonnes per hectare annually in 1985 and 1986 respectively. This action was intended to saturate the soil with the divalent cation Ca^{2+} in the soil exchange sites previously occupied by the monovalent Na^+ and H^+ ions. Ca^{2+} increases the flocculation ability of the soil particles, and is applied as a remedy for dispersed saline soil.

The maximum Magnesium flux (figure 4.16) is 1.4 mg/week, while the greatest Ca^{2+} flux is 3.9 mg/week. Figure 4.12 shows the similarity in variation between these divalent cations, although the concentrations of Calcium are higher than those of Magnesium. This accounts for the lower fluxes measured for Mg^{2+}. Potassium flux is low, except for 4 peaks, which do not correspond with any concentration peaks in figure 4.13, and are regarded as anomalous.

5. HYDROSALINITY MODELLING OF LEACHING FLUX WITHIN THE ROOT ZONE

Four hydrosalinity models are evaluated using field data representative of the lower Coerney River irrigation area. Three of these models were developed for management purposes and the fourth must be considered a research tool due to its intensive data requirements. The criteria for their selection is their variable levels of theoretical rigour and hence data requirements. These variations lead to differing levels in their ease of applicability for management purposes, which should be reflected by the accuracy of their results. The three management models examined in this study can be placed on a continuum of varying complexity from the very simple Leaching Requirement model (LR), to the SODICS model and the more complex PEAK model. The research model, LEACHC, is considerably more detailed and physically based than the selected management models. It has been suggested that the spatial and temporal accuracy of predictions should increase with increasing model complexity (Thorburn, 1988). However, to apply more complex physically based models, users require more intensive data which are not always readily available. Greater expertise is also required to run and interpret these more theoretical models. It is therefore important that model users select an appropriate model to meet their requirements. Detailed predictions of solute movement are not always required in irrigation management, and so the output generated by a simple capacity type model may be more appropriate than that from the more complex thermodynamic based models. Even with the use of field orientated models of low complexity there is a trade off between detail of prediction and data requirements. This situation must be resolved by the individual user, who must achieve a compromise between the required detail of prediction and practical constraints of data availability.

Detailed sensitivity analyses are presented for each of the management models, and the preliminary results of a sensitivity analysis carried out on the research model by Moolman (1992) are examined. The utility of each model is evaluated in terms of accuracy of prediction, ease of application and the costs involved in meeting the respective models data and computing requirements.

5.1 THE LEACHING REQUIREMENT MODEL

5.1.1 Model description

The concept of the Leaching Requirement (LR) model was developed by the U.S. Salinity Laboratory (Richards, 1954), and defines LR as the fraction of irrigation water that must drain through the root zone to keep the salinity of this zone below the limit that can be tolerated by a specific crop. The Leaching Requirement (LR) model is a simple lumped parameter representation of the mass balance. The amount and quality of applied irrigation water is assumed to determine the amount and quality of soil moisture flux leaving the root zone. The LR model does not account for the effects of soil physical properties, or any chemical interactions that may take place within the soil, thus steady state conditions are assumed.

The solute concentration of soil moisture in the root zone is determined by the balance between incoming solutes (Ci) and outgoing solutes (Cd). The LR model implies that by varying the amount of applied water it is possible to control the concentration of salts in the drainage water, and hence to maintain the concentration of the soil solution in the root zone at some level between the irrigation and drainage water concentrations. The relationship between volumes and salinities of the irrigation, soil and drainage waters can be expressed as:

$$Vd.Cd = Vi.Ci \quad (5.1)$$

where	Vd	=	volume of drainage water
	Vi	=	volume of irrigation water
	Cd	=	concentration of drainage water
	Ci	=	concentration of irrigation water

Thus the Leaching Fraction (LF) = Vd/Vi = Ci/Cd.

The input requirements of the LR model are very simple and can readily be applied without extensive sampling or expensive laboratory analyses. The model can be calculated on a pocket calculator, without the necessity for computer time. Although simple in concept, the LR model can be applied in a number of ways:

i. Given the irrigation water concentration and the desired soil solution concentration, the amount of irrigation to be applied can be determined.

ii. Given the irrigation water concentration and actual soil solution concentration, the leaching flux necessary to reduce salt build-up can be calculated.

iii. If over-irrigation is suspected, the effects of reduction in irrigation water of known concentration on the concentration of soil solution can be assessed before any change in scheduling is attempted.

The simplicity and small data requirements of the LR model allow its application without the need for highly trained personnel, specialised equipment or detailed analyses. However, the theoretical simplicity of the model results in several limitations to its application. The LR model assumes steady state conditions and it cannot account for the distribution of salts within the root zone, and for spatial and temporal variation of root zone salinity. These phenomena are dependent on soil characteristics and irrigation management strategies such as the frequency and rate of application, not just the depth and salinity of the irrigation water applied.

Temporal variations in soil solute concentration occur at a range of scales from localised short term effects between irrigations to long term field-wide changes over several years or decades. It is to quantify and predict such short to long term, and localised to wide-spread changes in soil salinity that leaching models are often required. As the LR model is applicable only in situations where steady-state conditions are assumed to exist, or reached very rapidly, it cannot be used with confidence in the majority of field situations. Where steady state conditions have not been reached, or where the solute balance is changing with land use development and other human impacts, the model assumptions limit the application of the LR model.

Although the LR model is not intended for short term fluctuations in soil salinity (Jury et. al., 1978; Rose et. al., 1979) it has been successfully used to model soil solutes at the level of 4 to 6 irrigation episodes (Bernstein and Francois, 1973). Bernstein and Francois (1971) used this model to examine alfalfa yields which were found to be relatively unaffected when the soil water salinity was allowed to approach the limits of this crops salt tolerance. They showed that much greater efficiency in water use was possible without significant reductions in yield and conclude that the leaching requirements for alfalfa could be reduced to 25% of the level previously thought essential for this crop of low to moderate salt tolerance.

5.1.2. Assessment of the LR model using field data

To assess the LR model's ability to predict leaching flux, information collected for the micro-plot site DDM03 in orchard M of Daisy Dell farm is used (Chapter 4). The intensive soil moisture and solute data collected for this site enables a comparison to be made between rates of flux determined by the LR model and those determined by the three methods described in section 4.4. The model input requirements are very simple comprising water quality parameters for the irrigation water which was sampled from the supply canal, the weekly combined depth of irrigation and rain water applied, and weekly soil solute concentrations which were extracted using in situ soil moisture extraction tubes. All samples were analysed at the Institute for Water Research, Rhodes University.

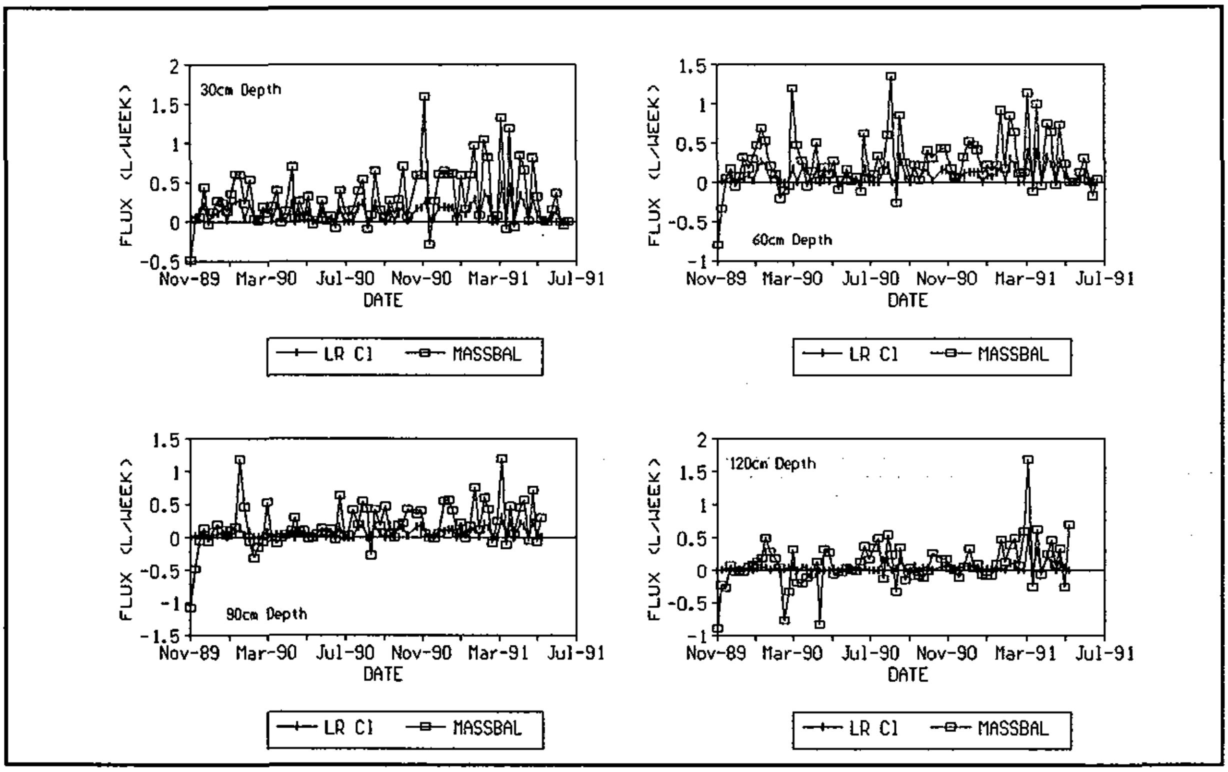


Figure 5.1. Comparison of weekly leaching flux determined using the LR model based on Chloride concentrations (LR Cl) and the Chloride mass balance approach (MASSBAL).

A spreadsheet package was used to apply the model. Concentrations of the major cations Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, and the Cl^{-} anion in the incoming irrigation water and the outgoing soil solution were entered to determine the leaching flux at 30, 60, 90 and 120 cm depths for 1 and 4 weekly intervals respectively. The LR model output are compared with the leaching flux values determined by the Chloride mass balance approach (tables 5.1 and 5.2) which was shown to provide the more realistic values of the three methods investigated in section 4.4.

Comparisons of the statistical summaries of leaching flux determined by the LR model, for both weekly and monthly intervals, with those determined using the Chloride mass balance approach shows the model to grossly under predict (table 5.1 and 5.2). Application of the LR model using the Chloride ion provides the best estimate, but even these values are at best only 30% of those determined using the Chloride mass balance approach. Comparison of the

Table 5.1 Comparison of weekly leaching flux values determined by the LR model and those determined by the Chloride mass balance approach.

		Maximum	Minimum	Mean	Std. Dev.
30 cm	Chloride	0.48	0	0.12	0.11
	Calcium	0.19	0	0.03	0.03
	Magnesium	0.30	0	0.05	0.06
	Sodium	0.28	0	0.07	0.07
	Potassium	0.04	0	0.01	0.01
	Cl Balance	1.59	-0.50	0.30	0.36
60 cm	Chloride	0.39	0	0.09	0.10
	Calcium	0.14	0	0.02	0.02
	Magnesium	0.21	0	0.04	0.04
	Sodium	0.35	0	0.05	0.06
	Potassium	0.02	0	0.00	0.00
	Cl Balance	1.35	-0.80	0.25	0.35
90 cm	Chloride	0.31	0	0.07	0.07
	Calcium	0.19	0	0.03	0.03
	Magnesium	0.20	0	0.05	0.05
	Sodium	0.15	0	0.03	0.03
	Potassium	0.03	0	0.01	0.01
	Cl Balance	1.19	-1.08	0.18	0.33
120 cm	Chloride	0.48	0	0.05	0.07
	Calcium	0.26	0	0.02	0.04
	Magnesium	0.28	0	0.04	0.05
	Sodium	0.13	0	0.02	0.03
	Potassium	0.09	0	0.02	0.02
	Cl Balance	1.69	-0.09	0.09	0.34

weekly and monthly modelled values show that similar mean rates of flux are predicted but with reduced maximum and increased minimum monthly values. These differences result from the averaging effect of the increased model time step. However, the overall results can only be described as disappointing and it is clear that all soil water drainage is not accounted for by the LR model. Conceptually, the LR model determines the Leaching Fraction as the ratio of the drainage and irrigated waters. It is clear that the ionic concentration of the solute extract sampled as being representative of the drainage water is too high. This may be due to the solute extract primarily sampling the micro pore flow and not the more rapid and in the DDM03 situation, more important macro pore flow which will be characterised by lower ionic concentrations. Owing to poor performance of the LR model and its inherent conceptual weakness due to major differences between the ionic concentration of macro and micro pore flow it was decided not to further examine the application of this model for predicting soil water drainage. Unfortunately this problem will be most prevalent in irrigated soils which are generally well draining.

5.2.4 Model evaluation

The LR model was developed as a tool for determining the leaching requirement of specific soils for the successful cultivation of particular crops using irrigation water of a given salinity. The model is extremely simple with very easily satisfied data requirements. However it was developed for steady state conditions with limited macro-pore flow. Although it has been widely used and proved a useful tool in some environments, especially those with a clay soil, it is not recommended for use within the Sundays River valley. The major

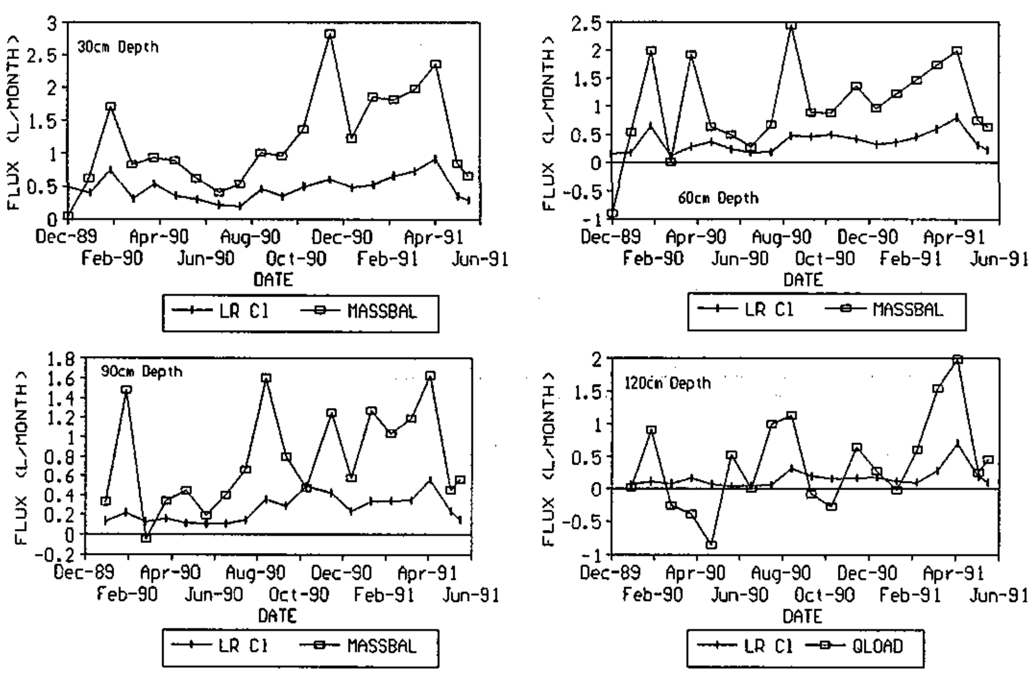


Figure 5.2. Monthly leaching flux determined using the LR model based on Chloride concentrations (LR Cl) and the Chloride mass balance approach (MASSBAL).

problem with this approach for determining the leaching flux is that the salinity of rapidly draining macro-pore flow is significantly less than the water in the soils micro-pores. This prevents one from simply applying the ratio of the extracted soil water and irrigation water salinities to determine the leaching flux. Therefore the Leaching Requirement model can not be recommended for use in free draining soils where macro-pore flow is dominant.

Table 5.2 Comparison of monthly leaching flux values determined using the LR model and those determined using the Chloride mass balance approach.

		Maximum	Minimum	Mean	Std. Dev.
30 cm	Chloride	0.92	0.20	0.47	0.18
	Calcium	0.31	0.03	0.10	0.07
	Magnesium	0.56	0.07	0.21	0.12
	Sodium	0.56	0.14	0.28	0.11
	Potassium	0.06	0.01	0.02	0.02
	Cl Balance	2.83	0.04	1.18	0.69
30 cm	Chloride	0.80	0.12	0.37	0.18
	Calcium	0.22	0.02	0.08	0.05
	Magnesium	0.36	0.05	0.14	0.08
	Sodium	0.49	0.07	0.21	0.12
	Potassium	0.03	0.01	0.01	0.01
	Cl Balance	2.44	-0.90	1.00	0.77
30 cm	Chloride	0.56	0.10	0.25	0.13
	Calcium	0.25	0.05	0.11	0.06
	Magnesium	0.40	0.10	0.19	0.08
	Sodium	0.30	0.04	0.10	0.06
	Potassium	0.06	0.02	0.03	0.01
	Cl Balance	1.62	-0.04	0.77	0.49
30 cm	Chloride	0.71	0.04	0.16	0.15
	Calcium	0.28	0.02	0.08	0.07
	Magnesium	0.40	0.03	0.14	0.09
	Sodium	0.30	0.02	0.06	0.06
	Potassium	0.14	0.01	0.07	0.04
	Cl Balance	2.00	-0.86	0.39	0.69

5.2 THE SODICS MODEL

The SODICS model is derived in a similar way to the Leaching Requirement (LR) model as it is also based on a simple mass balance of inputs and outputs within the soil. However, SODICS allows for changes in soil moisture and solute concentration through time by incorporating an empirically derived differential function. SODICS, developed by Rose et. al. (1979), uses Chloride as a tracer of soil moisture movement. The change in mean soil solute concentration (Cs) with time to a given soil depth is expressed according to the principle of conservation of mass:

$$Cz.\theta.\delta Cs/\delta t = ICi - LCz \qquad (5.2)$$

where Ci = solute concentration of the irrigation water (meq/l)
Cz = solute concentration of the soil at depth z (meq/l)
Cs = soil solute concentration
I = irrigation application rate (mm/year)
L = leaching rate at depth z (mm/year)
θ = soil water content at and above which leaching occurs (cm^3/cm^3)
t = time (years)
z = depth (mm)

To enable Equation 5.2 to be solved simply, a non-dimensional parameter P is defined as:

$$P = Cz/Cs \qquad (5.3)$$

Combining Equations 5.2 and 5.3 gives:

$$Cs_t = Cs_o + (I.Ci/LP)(1\text{-}exp^{(-L.P.t)/(z.\theta)}) \qquad (5.4)$$

where Cs_t and Cs_o = mean solute concentrations to depth z at times t and o.

A Newton-Raphson iterative process is used to determine the first approximation for L, by successively solving Equation 5.4. If downward leaching occurs, that is if L > 0, Cs increases with time to a final steady state value (Cf), which can be calculated from:

$$Cf = I.Ci/LP \qquad (5.5)$$

If L < 0, that is if upward leaching has occurred, Cs continually increases with time. Values of L and Cf may be used to assess the long term impact of an irrigation management strategy.

As the data requirements of this model are simple and easily satisfied it is applicable where more detailed information is not readily available. Values for I and Ci are obtained from information on the irrigation schedule applied. The soil parameters, such as bulk density, field capacity and the salinity and moisture profiles at two points in time may be determined from field surveys. The requirement for soil data at two points in time can be satisfied through substituting space for time and sampling an untreated and a treated site when the period of treatment is known (Rose, et. al., 1979).

The model determines soil Chloride concentrations for each time step until a steady state condition is reached. The model output consists of a table containing mean leaching flux, soil Chloride concentrations, steady state soil Chloride and the period of time required to reach steady state conditions.

5.2.2 Sensitivity analysis of the SODICS model

The SODICS model has 12 input parameters, each of which is varied in turn while all others are held at their mean values (table 5.3). The ranges over which the values are varied is based on the field data collected for the lower Coerney valley by the Institute for Water Research and the results of a study published by Folscher (1982). Therefore the data used in the sensitivity analysis are realistic extreme values for the soil types under the land use management found within the lower Coerney valley.

The sensitivity analysis examines the influence of the different model parameters in determining estimated values of the mean leaching flux, soil chloride profile and the period of time before steady state conditions are met. The leaching flux determined using the mean values for all model parameters is 213 mm/year. Under these conditions, the final Chloride concentration within the soil profile is 0.65 meq/l and the time taken to reach steady state concentration is less than one year.

Parameters influencing leaching flux

The effects of varying the input parameters for SODICS on the mean leaching flux predicted over the simulation period are summarised in table 5.4. Mean leaching flux under mean values for all parameters is calculated by SODICS to be 213 mm/year. Table 5.4 suggests that leaching flux predictions are most sensitive to variation of the Chloride concentration within the soil profile, where the maximum T_1 concentration produced a negative leaching flux of -14897 mm/year. The minimum soil solute concentration for the profile at depth z resulted in a positive leaching flux of 10638 mm/year, which is almost 10 times the combined annual total for both irrigation and rainfall (1300mm).

Air dry moisture (ADM) at T_o and T_1 also affected the leaching flux, but to a lesser extent than the profile Chloride concentrations. Maximum ADM at T_o and minimum ADM at T_1 both resulted in an increase in mean leaching flux, to 239mm/year and 279 mm/year respectively. Minimum ADM at T_o and maximum ADM at T_1 led to the opposite effect, with the mean leaching flux depressed to 184mm/year and 158 mm/year respectively. ADM at depth z had no effect on the leaching flux.

Field capacity (FC) had a greater effect on the mean leaching flux than variation of air dry moisture. The maximum FC resulted in a leaching flux of 355mm/year, with the minimum FC reducing the flux to 71mm/year. The only other parameter describing soil physical state is bulk density, which has no influence on the leaching flux.

Variation in the amount of irrigation applied was expected to have a major impact on the predicted mean leaching flux, but this is not shown by the sensitivity analysis (table 5.4).

Maximum irrigation resulted in a leaching flux of just over 100mm more than the mean of 213mm/year, while the minimum irrigation resulted in approximately 100mm less than the mean. Chloride concentration of the irrigation water influenced the leaching flux, but to a lesser extent than the irrigation amount. The maximum Cl concentration of 15 meq/l raised the leaching flux to 319 mm/year, while the minimum concentration of 5 meq/l led to a decrease in flux to 106 mm/year.

The most important parameter in determining the prediction of mean leaching flux by SODICS for this range of data is shown to be the profile soil Chloride concentrations.

Parameters influencing soil profile chloride concentrations

The influence of the different input parameters on the steady state Cl^- concentrations are shown in table 5.5. The concentration output is most sensitive to changes in the mean Cl^- concentration values of the soil profile at T_1 and T_o. The minimum profile concentration at T_o and the maximum concentration entered for time T_1 both result in an almost infinitely high solute concentration at steady state. Minimum concentration of T_1 results in a concentration of 0.013 meq/l, which is the lowest level obtained for any parameter variation. The maximum Cl^- concentration at T_o produced the second lowest steady state concentration of 0.157 meq/l. The only other parameter that impacts on the soil Cl^- concentration at steady state is the air dry moisture (ADM) content of the profile. Minimum ADM at T_o and all other moisture contents at T_1, depth z, had no effect on the steady state concentration which remained at 0.65 meq/l. Maximum ADM at T_o and minimum ADM at T_1 both led to a reduction in the steady state concentration, to 0.579 meq/l and 0.495 meq/l respectively. Maximum ADM at T_1 resulted in an increase in concentration to 0.876 meq/l. Variation in all other parameters had no effect on the predicted Cl^- concentration of the soil at steady state, apart from a slight decrease from 0.65 meq/l to 0.649 meq/l for the minimum field capacity.

Table 5.3 Range in parameter values for use in the sensitivity analysis of SODICS

MODEL PARAMETERS	PARAMETER VALUES		
	Maximum	Mean	Minimum
Rooting Depth (mm)	1200	600	200
Profile Cl at T_o	10.0	0.5	0.01
Profile Cl at T_1	10.0	0.5	0.01
Cl at T_1 and depth = z	10.0	0.5	0.01
Air Dry Moisture at T_o	50	30	10
Air Dry Moisture at T_1	50	30	10
Air Dry Moisture at T_1 and depth = z	50	30	10
Field Capacity	0.5	0.3	0.1
Bulk Density (g/cm^3)	2.0	1.5	1.0
Annual Irrigation (mm)	2000	1300	700
Irrigation concentration (meq/l)	15	10	5
Time Interval	100	20	1

Parameters influencing time to steady state

The concentration of Cl within the soil profile had the greatest impact on the predicted time to steady state salinity in the soil. Both the minimum T_o and maximum T_1 concentrations resulted in over 200 years before steady state is reached. The total number of time period iterations is limited to 200 years as the maximum length of time considered reasonable to model for the purpose of this study. The minimum time to steady state determined during the sensitivity analysis is 0 years, which is the time required to reach steady state when all parameters were set to their mean values. This is also obtained for a number of other input variations (table 5.6).

Air dry moisture (ADM) at T_o and T_1 all resulted in times to steady state of more than 10 years. However the ADM for depth z at T_1 had no effect on the time to reach steady state which remained at 0 years. Maximum field capacity (FC) had no impact on the time to steady state, although minimum FC increased the mean time to 3 years. Bulk density had no effect on the time taken. Irrigation amount and concentration had a slight impact on the time to steady state, with the minimum values leading to an increase of 2 and 3 years respectively over the mean time taken. The factor most influential in determining time to steady state is the soil profile Chloride concentration.

Overall the most influential input parameters affecting the predictions of leaching flux, steady state soil concentration, and time to reach steady state were the soil solute concentration values. Soil physical properties were less important in determining output variability, for example the bulk density had no effect on the model output at any value. The crop related input parameters consisting of the rooting depth, also had no effect on model performance. It is concluded therefore that the accuracy in field measurement of the soil Chlorides is the most important factor in obtaining the input requirements necessary to run this model. Focus on the soil physical properties need not be as detailed as the soil chemistry.

Table 5.4 Variation in parameters and mean leaching flux predicted by SODICS.

PARAMETER VALUES	Maximum parameter value		Minimum parameter value	
	Parameter Value	Leaching flux (mm/year)	Parameter Value	Leaching flux (mm/year)
Rooting Depth (mm)	1200	L= 213	200	L=213
Profile Cl at T_o	10.0	L= 880	0.01	L=-109
Profile Cl at T_1	10.0	L=-14897	0.01	L=213
Cl at T_1 and depth = z	10.0	L= 10.64	0.01	L=10638
Air Dry Moisture at T_o	50	L = 239	10	L=184
Air Dry Moisture at T_1	50	L = 158	10	L=279
Air Dry Moisture at T_1 and depth=z	50	L=213	10	L=213
Field Capacity	0.5	L=355	0.1	L=71
Bulk Density (g/cm^3)	2.0	L=213	1.0	L=213
Annual Irrigation (mm)	2000	L=327	700	L=115
Irrigation concentration (meq/l)	15	L=319	5	L=106
Time Interval	100	L=213	1	L=213

Table 5.5 SODICS predictions of soil Chloride concentration at steady state.

MODEL PARAMETERS	Maximum Parameter Values		Minimum Parameter Values	
	Parameter Value	Cl (meq/l)	Parameter Value	Cl (meq/l)
Rooting Depth (mm)	1200	c=0.65	200	c=0.65
Profile Cl at T_0	10.0	c=0.157	.01	c=****
Profile Cl at T_1	10.0	c=****	.01	c=0.013
Cl at T_1 and depth = z	10.0	c=0.649	.01	c=0.65
Air Dry Moisture at T_0	50	c=0.579	10	c=0.65
Air Dry Moisture at T_1	50	c=0.876	10	c=0.495
Air Dry Moisture at T_1 and depth = z	50	c=0.65	10	c=0.65
Field Capacity	0.5	c=0.65	0.1	c=0.649
Bulk Density (g/cm^3)	2.0	c=0.65	1.0	c=0.65
Annual Irrigation (mm)	2000	c=0.65	700	c=0.65
Irrigation concentration (meq/l)	15	c=0.65	5	c=0.65
Time Interval	100	c=0.65	1	c=0.65

Table 5.6 SODICS predictions of time to steady state.

Model Parameters	Maximum Parameter Values		Minimum Parameter Values	
	Parameter Value	Time to Steady State (years)	Parameter Value	Time to Steady State (years)
Rooting Depth (mm)	1200	Y=1	200	Y=0
Profile Cl at T_0	10.0	Y=5	0.01	Y=>200
Profile Cl at T_1	10.0	Y=>200	0.01	Y=1
Cl at T_1 and depth = z	10.0	Y=4	0.01	Y=1
Air Dry Moisture at T_0	50	Y=11	10	Y=15
Air Dry Moisture at T_1	50	Y=18	10	Y=10
Air Dry Moisture at T_1 and depth = z	50	Y=0	10	Y=0
Field Capacity	0.5	Y=0	0.1	Y=3
Bulk Density (g/cm^3)	2.0	Y=0	1.0	Y=0
Annual Irrigation (mm)	2000	Y=0	700	Y=2
Irrigation concentration (meq/l)	15	Y=0	5	Y=3
Time Interval	100	Y=0	1	Y=0

Table 5.7 Weekly leaching fluxes (mm/week) determined using the Sodics model and the three methods discussed in section 4.2.

		Maximum	Minimum	Mean	Std. Dev.
30 cm	SODICS	193.37	-82.92	4.01	40.16
	MASSBAL	70.71	-49.99	21.66	24.51
	MATRIC	14.83	-18.59	0.11	6.83
	WATBAL	80.85	-191.77	2.63	38.68
60 cm	SODICS	205.10	-85.02	3.40	43.90
	MASSBAL	135.37	-79.91	25.77	35.55
	MATRIC	32.83	-34.89	-0.12	13.63
	WATBAL	137.38	-366.07	8.51	66.10
90 cm	SODICS	152.00	-74.31	7.15	42.85
	MASSBAL	118.64	-48.37	15.57	26.30
	MATRIC	22.40	-30.96	0.18	10.61
	WATBAL	112.15	-76.00	8.15	35.16
120 cm	SODICS	148.78	-69.57	4.75	44.79
	MASSBAL	53.77	-89.55	3.61	26.59
	MATRIC	15.36	-21.05	0.07	6.45
	WATBAL	135.59	-719.43	-6.07	107.47

where MATRIC = Matric Potential Flux measurement WATBAL = Water Balance Flux measurement
MASSBAL = Volumetric Mass Balance Flux measurements

Table 5.8 Monthly leaching fluxes (mm/month) determined using the Sodics model and the three methods discussed in section 4.2.

Method of Estimation		MAXIMUM	MINIMUM	MEAN	STD.DEV.
30 cm	SODICS	177.12	-82.34	11.09	68.64
	MASSBAL	204.67	3.79	111.68	58.53
	MATRIC	18.12	-17.89	0.22	10.50
	WATBAL	185.31	-161.74	36.81	70.33
60 cm	SODICS	185.81	-66.02	24.61	62.64
	MASSBAL	244.36	-89.57	98.02	81.44
	MATRIC	30.53	-34.89	0.81	17.50
	WATBAL	178.68	-215.76	34.29	79.45
90 cm	SODICS	99.53	-97.60	12.04	60.97
	MASSBAL	160.05	-149.09	61.73	70.28
	MATRIC	27.01	-30.30	0.41	15.60
	WATBAL	158.77	-224.47	32.77	76.67
120 cm	SODICS	241.94	-58.47	66.12	100.30
	MASSBAL	153.91	-131.34	22.04	73.33
	MATRIC	13.65	-19.34	0.01	9.23
	WATBAL	147.70	-230.02	32.87	80.17

5.2.3 Assessment of the SODICS model using field data

To validate the Sodics model, information collected at the micro-plot site DDM03 on Daisy Dell farm is used to predict the mean leaching flux within the soil profile at depths of 30, 60, 90 and 120 cm for time steps of 1 and 4 weeks respectively. Thorburn et. al. (1987)states that this model can be applied over any time increment, providing that sufficient change has occurred in the soil salinity to allow the calculation of the decay curve, equation 5.4. Statistical summaries of the output from these model runs are compared with the leaching flux values estimated using the three methods examined in Chapter 4.

Clearly the leaching flux values determined to a depth of 90 cm using the sodics model grossly underpredict those determined by the Chloride mass balance approach. This approach was found to be the most appropriate for estimating leaching flux at site DDM03 where macropore flow is an important component of the soil drainage process. Below 90 cm the mass balance approach is thought to underestimate leaching flux due to the increased lateral dispersion of solute carrying soil moisture. Therefore the comparison of flux below this depth is tenuous. The sensitivity analyses show that the sodics model is most sensitive to the soil profile chloride concentration parameters at T_o and T_1. The values of these parameters determine the solution of Eq. 5.4 which controls the rate of change of the chloride concentration within the soil profile which indirectly determines the predicted soil moisture drainage. The sensitivity analyses also show that if the Chloride concentration at T_t is too high then the model will under predict leaching. Therefore it seems that as macro-pore flow is an important component of the drainage process at DDM03, the Cl concentration of the soil extracts collected for DDM03 are greater than the mean concentration of the water leaching through these soils. With such incorrect parameter values, Eq. 5.5 is unable to properly describe variations in the chloride concentration and the model is unable to properly model leaching flux.

5.3 THE PEAK MODEL

5.3.1 Model Description

The PEAK model is based on the convective-dispersion equation for solute and moisture movement through porous media under transient conditions. It is of a similar order of simplicity to the steady state SODICS model, but has been demonstrated to give better predictions of recharge rates (Thorburn, et. al., 1991).

The movement of a peak of non-adsorbed, non-reactive solute through a soil under constant moisture flux was modelled by De Smedt and Wierenga (1978). The peak was modelled in two parts: the first section predicts the movement of a solute peak through a soil profile (PEAKM), and the second predicts the dispersion of the solute about the peak (PEAKD).

PEAKM

The original requirement for constant moisture flux (De Smedt and Wierenga, 1978) was alleviated by Rose et. al (1982) using a mass balance approach. The movement of the peak solute concentration through uniform soil is described by:

$$J = qC - Ds.\theta.\delta/\delta z \quad \ldots\ldots \quad (5.6)$$

where J = solute flux
q = volumetric flow rate
Ds = diffusion coefficient in soil
θ = soil water content
C = solute concentration
z = soil depth

The model assumes:

i. there is no preferential movement of water through the soil, such as macro-pore flow, lateral flow, soil cracking or other instability effects (Thorburn, 1988).

ii. the soil drains to a constant water content after each infiltration. This is termed field capacity.

iii. no significant soil-solute interactions or chemical transformations occur.

iv. crop water uptake is uniform with depth to the base of the root zone.

v. the profile is at field capacity below the root zone.

Equation 5.6 was then modified by Rose et. al., (1982) to account for plant solute uptake:

$$\theta\delta C/\delta t = C.\delta q_r/\delta z - q\delta C/\delta z \quad \ldots\ldots \quad (5.7)$$

where q_r = volumetric water uptake by plant roots.
q = volumetric flow rate
θ = soil water content
C = solute concentration

Water input to and loss from the soil profile is considered separately as an infiltration event and an evapotranspiration event respectively. During infiltration, q_r is much less than the water flux, so Equation 5.7 reduces to Equation 5.6. During evapotranspiration, the reverse occurs, with q_r much greater than the water flux which is approximately zero. In this phase there is a negligible change in the position of the concentration peak but an overall increase in solute concentration.

The depth beneath the soil surface of the solute concentration peak is defined as the depth at which the solute flux, $\delta C/\delta z$, = 0. This equation neglects the effect of dispersion, which has been shown to result in an overestimation of approximately 10% of the depth predicted (Rose et. al., 1982). This error is considered acceptable within the context of the model application.

During an infiltration event the depth of a solute peak (a) can be represented by:

$$\delta a/\delta t = q/\theta_{(z=a)} \quad (5.8)$$

where	a	=	depth beneath the soil surface of the solute peak
	z	=	soil depth
	q	=	volumetric flow rate
	θ	=	soil moisture content

During an evapotranspiration event $\delta a/\delta t = 0$. Equation 5.8 has been shown to generate an acceptable approximation of solute peak movement in laboratory studies (Warwick et. al., 1971; Ghuman et. al., 1975; De Smedt and Wierenga, 1978).

Equation 5.8 is then presented in a discontinuous form. The solute peak movement within a profile is calculated by assuming that the solute is displaced by the wetting front of successive water applications, between which evapotranspiration occurs (Rose et. al., 1982). The depth below the soil surface of the solute peak is calculated in finite time intervals:

$$a_n = a_{(n-1)}/\theta_{fc} + \Sigma_W \quad (5.9)$$

where	a	=	depth of solute peak
	n	=	number of time element
	θ_{fc}	=	soil moisture content at field capacity
	Σ_W	=	equivalent ponded depth of water which passed below previous peak position

A general expression for Σ_W is:

$$\Sigma_W = I.n.t - g(\theta_{fc} - \theta_{(n-1)} + j.E_t.t/r) \quad (5.10)$$

where	I	=	infiltration rate over the period of simulation
	n	=	number of event
	t	=	time interval
	g	=	depth of water required to bring soil moisture to field capacity
	θ_{fc}	=	soil moisture at field capacity
	$\theta_{(n-1)}$	=	soil moisture content at the end of the previous time increment
	j	=	proportion of evapotranspiration occurring before infiltration
	E_t	=	evapotranspiration rate over the period
	r	=	depth of rooting

If peak movement occurs, two extreme boundary condition are possible: either infiltration is complete before evapotranspiration begins (j = 1), or evapotranspiration ends before infiltration begins (j = 0). If values for j cannot be determined with confidence from irrigation scheduling data, the extreme values can be used to define envelope curves encompassing the uncertainty in prediction of peak movement.

The value of g is constrained by the previous depth of the peak and the rooting depth. If the previous depth to peak lies above the rooting depth, g is equal to the previous depth. Conversely, if the previous depth to peak lies below the rooting depth then g is taken as equal to the rooting depth.

The previous profile water content must be known to determine $\theta_{(n-1)}$, which is calculated by:

$$\theta_{(n-1)} = \theta_{(n-2)} + (I_{(n-1)} - E_{(n-1)}.\delta r/\delta t) \quad (5.11)$$

Soil moisture content during the previous time period ($\theta_{(n-1)}$) is constrained by the field capacity of the soil considered and the wilting point of the crop grown. If $\theta_{(n-1)}$ is greater than the field capacity then the soil moisture content is assumed to be equal to field capacity. Conversely, if $\theta_{(n-1)}$ is less than wilting point, then the soil moisture is assumed to be equal to the wilting point.

The data required to run the PEAKM model comprises information on the irrigation application, rainfall, evapotranspiration and crop rooting depth.

PEAKD

The convective dispersion equation based on the mass conservation of a non-adsorbed, non reactive solute requires that:

$$\delta.\theta C/\delta t = -\delta J/\delta z \quad (5.12)$$

where θC = mass of solute per unit soil volume
J = flux of solute
z = depth

Mass movement of water will give a convective flux of solute (Jc) which can also be described as qC. Water movement will also result in a diffusion like flux of solute (Jd) proportional to the concentration gradient. According to Fick's first law:

$$Jd = -Ds.\theta.\delta C/\delta z \quad (5.13)$$

where Ds = diffusion coefficient in soil
θ = soil moisture content
C = concentration of solute
z = depth

The total contribution of both convective (Jc) and diffusive (Jd) fluxes constitutes the convective dispersion equation:

$$J = Jc + Jd \quad (5.14)$$
$$= qC - Ds.\theta.\delta C/\delta z$$

Substituting Equation 5.13 into Equation 5.12 gives:

$$\delta C/\delta t = D.\delta^2 C/\delta z^2 - (q/\theta).\delta C/\delta z \quad (5.15)$$

$$= D.\delta^2 C/\delta z^2 - v.\delta C/\delta z \quad (5.16)$$

where $q/\theta = v$ = pore water velocity

De Smedt and Wierenga (1978) solved Equation 5.15 for constant flux boundary conditions, using values for the depth of the solute peak (a) from Equation 5.9:

$$C/C_o = 0.5 \text{ erfc}[(z-a)/2(Ds.t + a.E)^{0.5}] \quad (5.17)$$

where C_o = initial solute concentration
C = solute concentration
erfc = error function term (Carslaw and Jaeger, 1959)
z = depth of soil
a = depth of solute peak below surface
Ds = molecular diffusion coefficient in soil
E = solute dispersivity coefficient

Equation 5.16 calculates the leading edge of the dispersed solute pulse. The trailing edge (b) is given by:

$$b = a - L \quad (5.18)$$

where b = depth of the step decrease in solute concentration (trailing edge)
a = depth of solute peak below surface
L = depth of the solute pulse (leading edge)

When b is used to solve Equation 5.17:

$$C/C_o = 0.5 \text{ erfc}[(z-b)/2(Ds.t + b.E)^{0.5}] \quad (5.19)$$

As the solute pulse depth is given by a - b, Equation 5.19 is taken from Equation 5.17:

$$C/C_o = 0.5 \text{ erfc}[(z-a)/2(Ds.t + a.E)^{0.5}] - 0.5 \text{ erfc}[(z-b)/2(Ds.t + b.E)^{0.5}] \quad (5.20)$$

Values for L, E and Ds are required to solve Equation 5.20. Where the solute applied is through the irrigation water:

$$L = I.t/\theta_{fc} \quad (5.21)$$

where I = infiltration rate
t = time
θ_{fc} = field capacity

E must be determined through calibration of predicted output with measured data. Ds is taken as approximately 60mm²/day (Rowell et. al., 1967).

5.3.2 Sensitivity analysis of the PEAK model

The PEAK model is divided into two parts, the first (PEAKM) predicts the movement of a solute peak through the soil and the second (PEAKD) uses the output from PEAKM to predict the dispersion of a solute about the peak. These modules will be analysed separately for the purpose of determining the sensitivity of model output to variation in input parameters.

Sensitivity Analysis of PEAKM

This model has six input parameters, each of which was varied in turn between the extremes used in the SODICS sensitivity analysis (table 5.9). The effect of variation in each parameter on the model output is shown in the following graphs.

Depth of infiltrating water (irrigation and rainfall): (I)

This parameter was varied between the extremes of 250 and 50mm per month (figure 5.3). The maximum and minimum values selected for I were compiled from the irrigation scheduling data and measured rainfall. Only when I exceeds E can movement occur, which is whenever I is equal to, or greater than 70mm. Therefore at the minimum I value of 50mm, no movement occurs. When the mean value of I is used, the modelled peak moves to a depth of 4000mm over 12 time intervals, and the maximum I results in a depth of 10,000mm.

Table 5.9 Parameter values used for the sensitivity analysis of the PEAKM model

	Maximum	Mean	Minimum
I (rainfall + irrigation mm/month)	250	100	50
E (evaporation mm/month)	300	130	30
Pan Factor	0.95	0.7	0.45
Rooting Depth (mm)	1200	600	200
J (proportion of evapotranspiration occurring before infiltration)	1.0	0.5	0.0
Field Capacity (mm^3/mm^3)	0.5	0.3	0.1

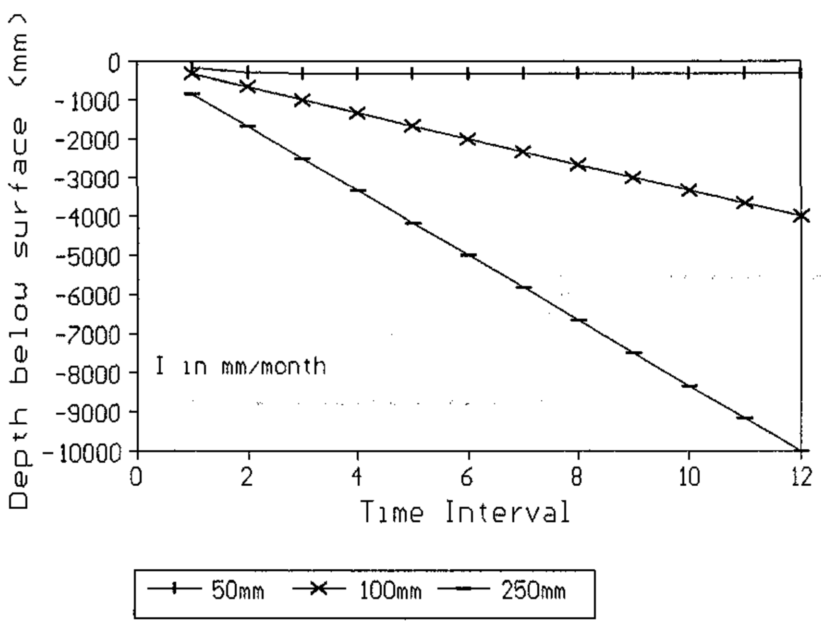


Figure 5.3 Effects of variation in irrigation and rainfall (I) on the peak depth predicted using PEAKM.

Depth of evaporation: (E)

The original model listings were modified to allow pan evaporation and a crop factor to be used, instead of a combined value for E which is difficult to quantify. E was varied between 30mm and 300mm/month (figure 5.4), with a mean of 130mm/month, and the crop factor held constant at 0.7. The extremes for E were derived from evaporation data obtained from the Department of Agriculture Citrus Research Station at Addo, located approximately 4km from the micro-plot on Daisy Dell farm.

As with I, the E parameter is instrumental in determining peak movement. E interacts with I to control the amount of water available for downward movement within the soil. The maximum E value of 300mm/month produced no peak movement at all over the simulation period. This is because maximum E exceeded the mean I of 100mm/month. The peak would not be expected to move unless actual evapotranspiration (E x Pan Factor) is less than the depth of irrigation and rainfall (I).

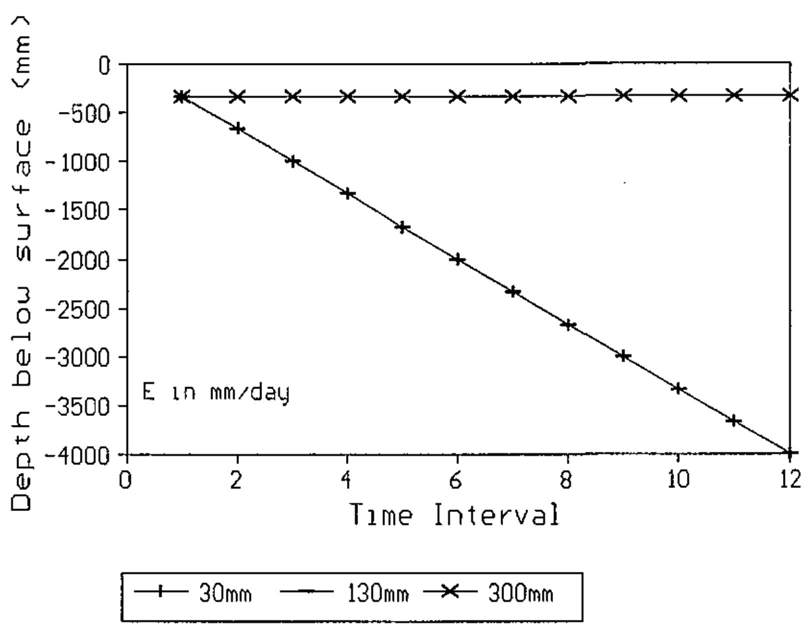


Figure 5.4 Effects of variation in depth of evaporation (E) on the peak depth predicted using PEAKM.

The mean E of 130 mm/month is greater than mean I of 100mm/month, but peak movement occurs due to the reduction in E through the use of the pan factor (the mean value of which is 0.7). The pan factor reduces the evaporation and only when E = 143 mm/month will peak movement be zero. Any value of E below this results in peak movement with no variation in rate of movement. Once E falls below 143mm/month the peak produced is identical to that produced using total mean conditions. The parameter E affects model output in an all or nothing manner, with no movement at all above E = 143 mm/month, and a constant movement rate and depth below E = 143 mm/month.

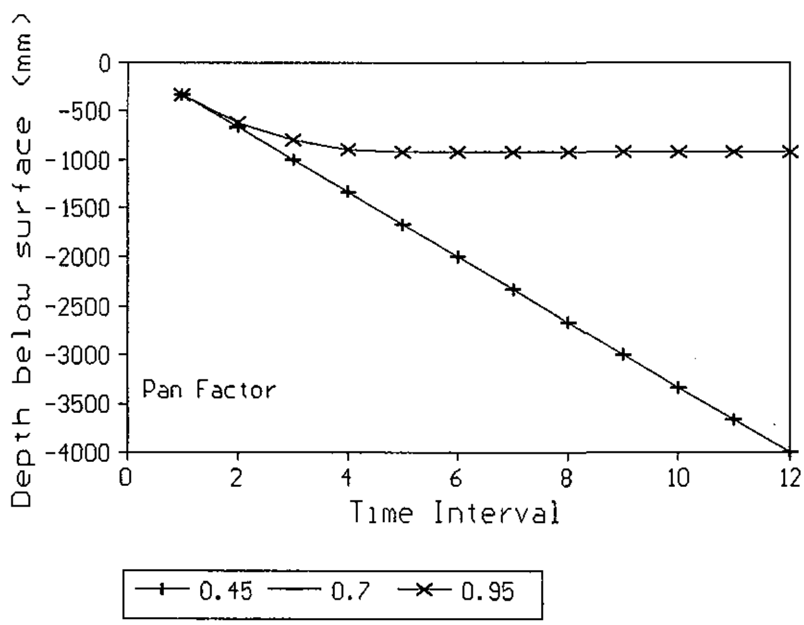


Figure 5.5. Effects of variation in the pan factor on the peak depth predicted using PEAKM

Pan Factor: (PF)

This parameter was varied between 0.95 and 0.01, with a mean of 0.7 as shown in figure 5.5. All values less than 0.7 produced peak movement equivalent to that of total mean conditions. Values greater than 0.7 result in a systematic decrease in peak movement, which is to be expected: as the PF value approaches one, the model output will approach the straight line, with no change in depth, as was calculated when E exceeds 143mm/month.

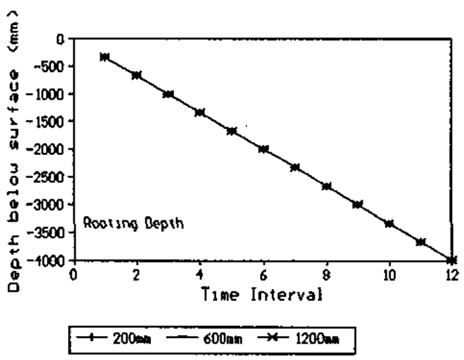


Figure 5.6 Effect of variation in rooting depth on predicted peak depth.

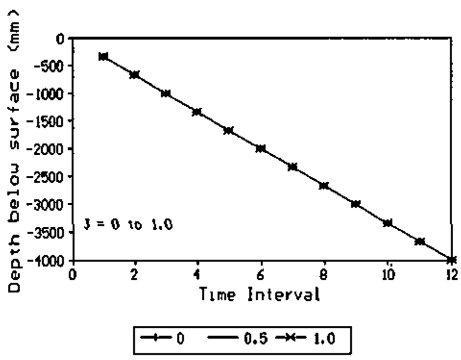


Figure 5.7 Effect of variation in J on on predicted peak depth.

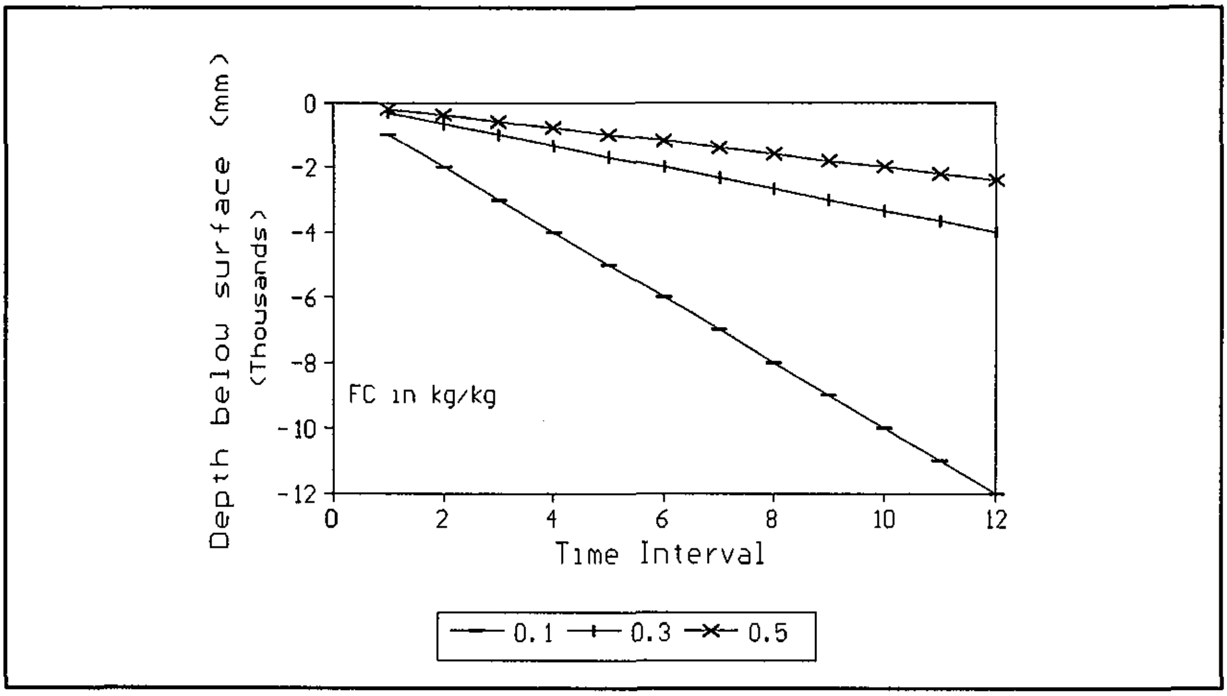


Figure 5.8 Effect of variation in field capacity on PEAKM model output.

Rooting Depth

Maximum and minimum rooting depths of 1200mm and 200mm respectively were taken from Folscher's (1982) extensive survey of the entire Sundays River valley. The mean rooting depth of 600mm was the depth determined by inspection of the orchard sampled. The rooting depth had no effect whatsoever on the peak movement predicted by the model (figure 5.6).

Proportion of evaporation occurring before infiltration: (J)

The proportion of evaporation occurring before infiltration (J) is a subjective parameter, requiring interpretation of the irrigation management data. The value of J lies between 0 and 1 and must be estimated from irrigation scheduling information, or calibration of the model against measured data. As this parameter had no effect on model output in this study, it was not considered an important model parameter (figure 5.7).

Field Capacity: (FC)

This parameter was varied between 0.5 and 0.1, which are the extreme values for a clay-rich soil and a sand respectively (figure 5.8). Of the soil physical properties accounted for in this model, FC is shown to have the most impact on the output peak movement. Although difficult to quantify in the field, FC was taken to be equivalent to the soil moisture content measured after saturation (matric potential = 0kPa) was reached during a three day storm event. FC down the study profile varied between 0.28 and 0.35, with 0.3 taken as the mean. Peak movement occurring when FC equalled 0.1 was approximately 2 500mm, and the mean FC produced a movement to 4 000mm, an increase of only 1 500mm. However between 0.3 and 0.5 the distance travelled by the peak was 8 000mm, which illustrates that at high field capacities, the effect of this parameter on model output becomes more crucial than at low field capacities.

5.3.3 Sensitivity analysis of PEAKD

The PEAKD model requires an input table generated by PEAKM, as well as other parameters. As the most influential soil hydraulic parameter affecting PEAKM output was the field capacity, the sensitivity analysis on PEAKD was performed using the maximum, mean and minimum field capacity output tables from PEAKM. The other parameters are

Table 5.10 Variation in parameters for the sensitivity analysis of the PEAKD model

	Maximum M	Mean	Minimum 1	Minimum 2
PEAKM INPUT using variable FC	0.5	0.3	0.1	
Diffusion coefficient (mm^2/month)	300	100	10	
Dispersion coefficient(mm)	150	100	50	5
Initial depth of infiltrating water(mm)	27	9	1	
Irrigation concentration (meq/l)	15	10	5	

summarised in table 5.10. Neither the maximum calculated depth, nor the depth increment are intended to have any effect on the model output. The depth increment allows greater detail to be obtained for smaller depths, but for the purpose of the sensitivity analysis, this was unnecessary.

Diffusion Coefficient: (D0)

Variation in this parameter had no effect on the model performance (figure 5.9). Output variations were due to differences in the input from PEAKM. The input table resulted in altering the output concentration curve slightly between the maximum FC and mean FC, with peak concentrations of 0.0555 to 0.0557 meq/l for the maximum and mean FC respectively. The minimum FC PEAKM input table produced more noticeable results altering the shape of the concentration curve, and elevating the peak concentration to 0.065 meq/l.

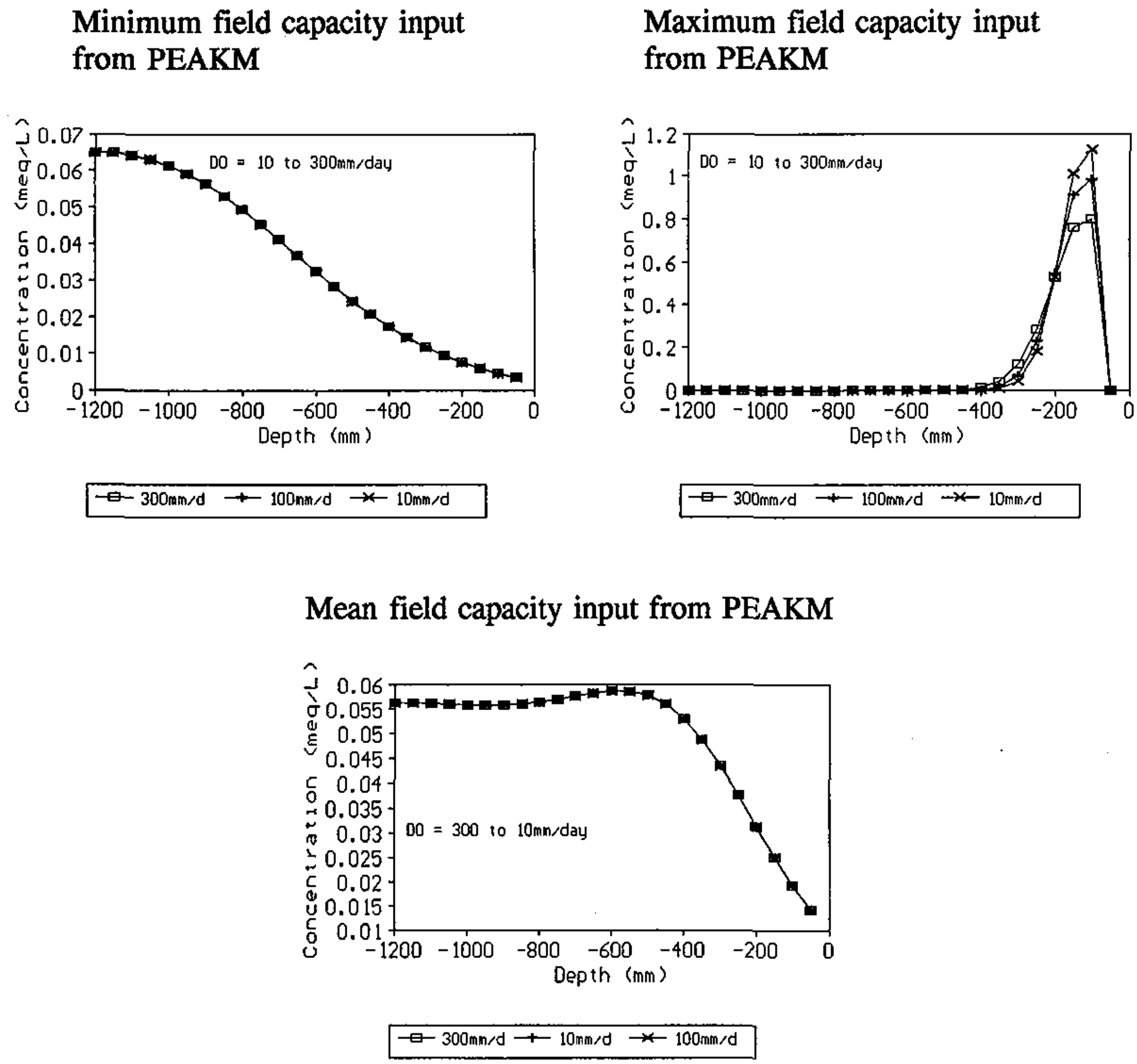


Figure 5.9 The effects of variation in diffusion coefficient on predicted peak depth and concentration using maximum, minimum and mean field capacity to determine PEAKM inputs.

Dispersion Coefficient: (DO)

Variation in the dispersion coefficient (D) for the range of values published in the literature (Rowell, et. al., 1967; Barry, et. al., 1985; Cameron and Wild, 1982; Rose, et. al., 1982) of 30mm to 120mm produced extremely variable model results. Incremental values for D were then entered to determine the direction and magnitude of response in output resulting from smaller changes in D. Values adopted were from 10mm at intervals of 10 to 100mm, with a minimum of 5mm and maximum of 150mm (figure 5.10). The impact of variations in low levels of D, ranging from 5 to 50 mm is high, but D becomes less effective between 60 and 150mm in influencing the peak curve. For sandy clays and clay rich soils, Rowell et al (1985) suggest that dispersion coefficients of between 60 an 100 mm should be used for determining the movement of Chlorides. A dispersion coefficient value of 60mm was adopted for this study.

When the maximum field capacity (FC) input from PEAKM was used, the final concentration curve was less variable than with the mean FC input, and the range in concentrations of the peak reduced from 0.12 meq/l at 300mm depth below surface to 0.1 meq/l at 200mm. However, the minimum FC table produced a concentration curve unlike either the mean or maximum FC. Minimum FC resulted in peak concentrations generally higher than that for the mean or maximum FC, at around 0.22 meq/l, and the depth of peak is much greater, at 1000mm. Therefore at low FC values, choice of D is more influential in determining the output from PEAKD than at intermediate or high values of FC.

Initial depth of infiltrating water: (IO)

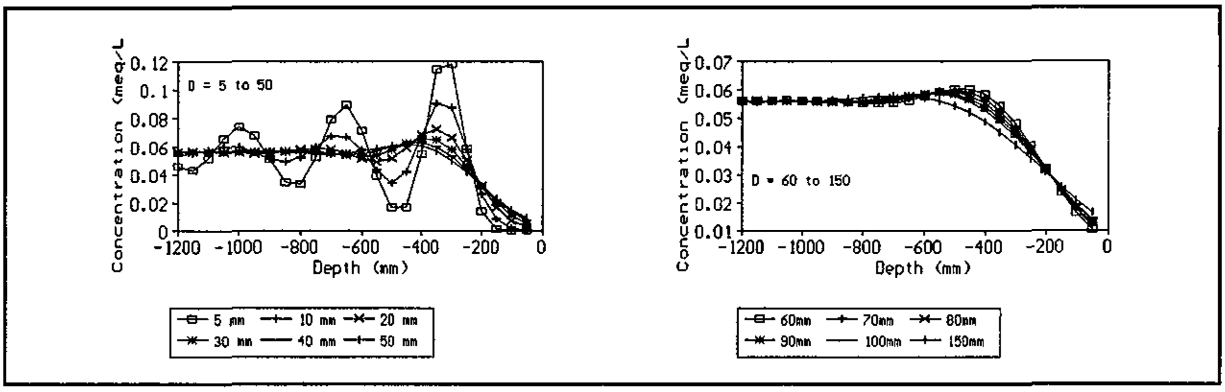
Given a field capacity of 0.3, the depth of moisture required to raise the soil moisture to field capacity within a 300mm layer was calculated to be 90mm. IO was then varied by factors of 9, so that the initial conditions at depths of 300, 600, 900 and 1200 mm were at field capacity. The concentration curves generated in this way using the PEAKM input table with the mean FC value have a similar shape, with the peak also occurring at a depth of 450mm, but with all concentrations displaced by approximately 0.05 meq/l for the peak concentration (figure 5.11). The maximum FC value produces almost identical peak concentration curves, but with the peak slightly shallower, at around 350mm. For the minimum FC input, peak concentrations are higher at all IO values, and the depth to the peak is greater, at around 1000mm.

Irrigation concentration: (CO)

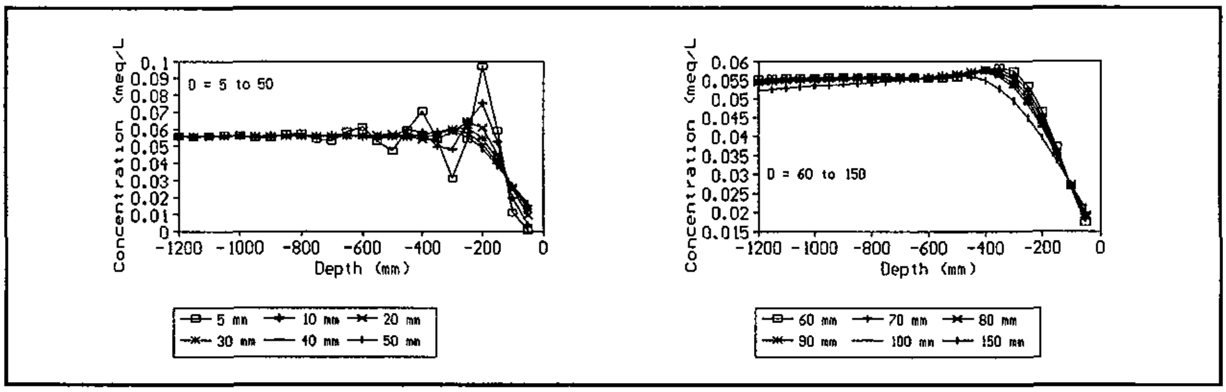
Figure 5.12 shows that this is an important input variable in determining the concentration of the peak. The shape of the soil concentration curve follows a similar trend for all PEAKM inputs and all irrigation concentrations (CO). However, the concentrations of the solute peaks vary between 5 and 0 meq/l.

Peak concentrations vary between 0.09 and 0.03 meq/l for the irrigation concentrations (IO) of 15 and 5 meq/l respectively when mean field capacity (FC) input from PEAKM is used. The depth of peak is similar for all concentrations of irrigation water, occurring at 550mm depth. The maximum FC input results in peak concentrations slightly lower than the mean FC input, ranging from 0.085 to 0.028 meq/l for the maximum and minimum concentrations

Mean Field Capacity PEAKM input



Maximum Field Capacity PEAKM inputs



Minimum Field Capacity PEAKM input

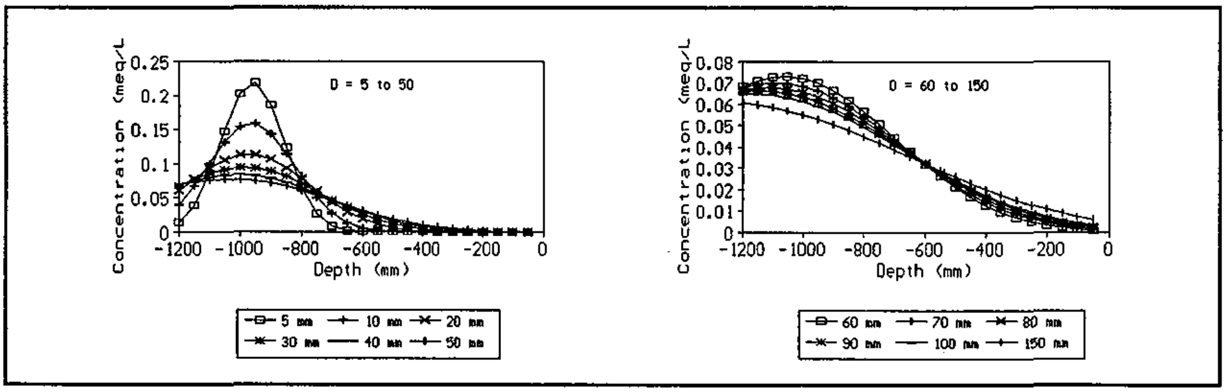
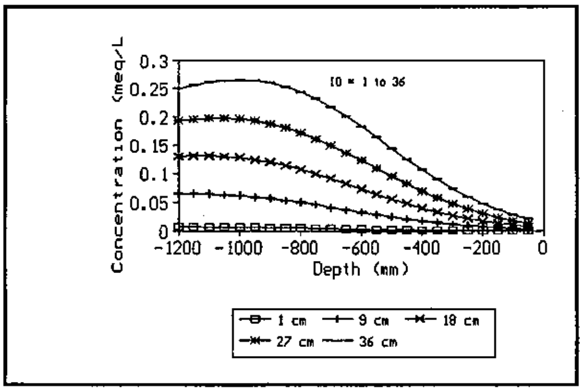
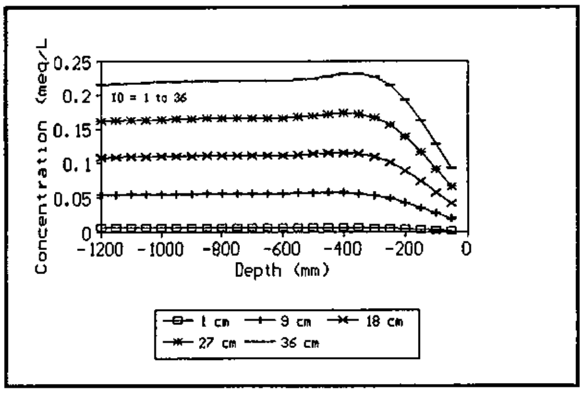


Figure 5.10 Effect of variation in Dispersion Coefficient on predicted peak depth and concentration, using maximum, minimum and mean Field Capacity to determine PEAKM inputs

Minimum field capacity input from PEAKM



Maximum field capacity input from PEAKM



Mean field capacity input from PEAKM

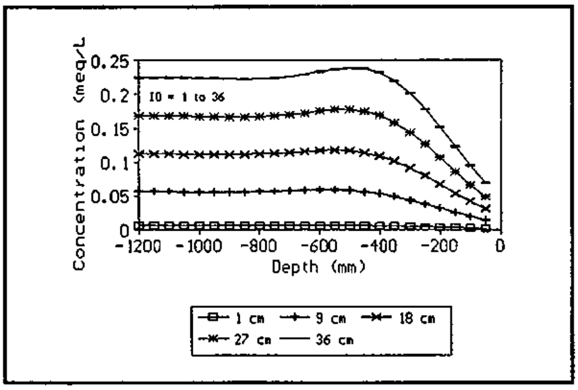


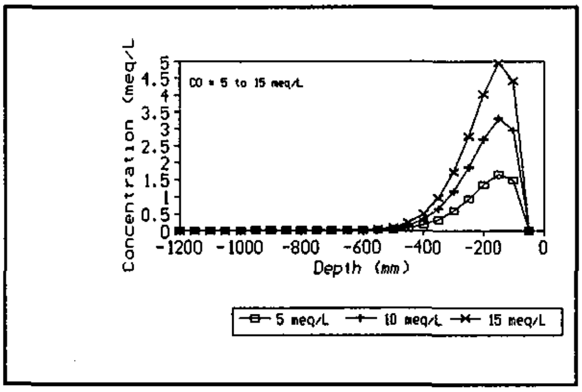
Figure 5.11 Effect of variation in initial profile moisture on peak depth and concentration, using maximum, minimum and mean field capacity to determine PEAKM input.

respectively. The peak depth is slightly shallower, at around 425mm. Minimum FC input produces a concentration peak curve, with higher concentrations, ranging from 0.098 to 0.033, and a deeper peak at around 1200mm.

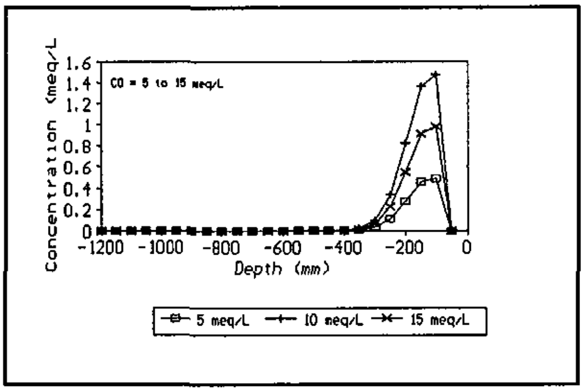
5.3.4 Assessment of the PEAK model using field data

The PEAK model has been successfully applied at daily (Barry, et. al., 1985), monthly (Rose, et. al., 1982), and annual (Thorburn, 1988) time intervals. For this assessment of the models ability to predict irrigation drainage it is applied at both weekly and monthly time steps. Soil moisture and soil moisture flux are modelled at the 120 cm depth of the soil profile at micro-plot site DDM03 in orchard M of Daisy Dell farm. These soil moisture and flux values are compared to those determined using the matric potential and soil water solute data collected at this site.

Minimum field capacity input from PEAKM



Maximum field capacity input from PEAKM



Mean field capacity input from PEAKM

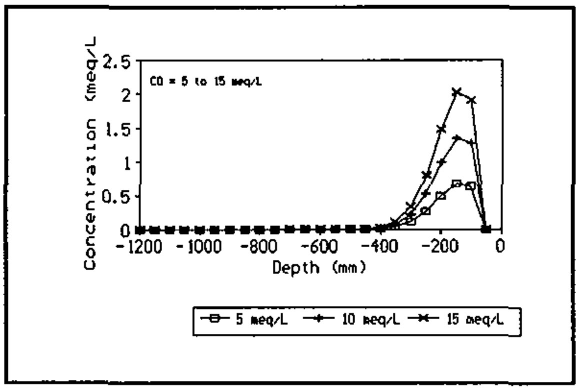


Figure 5.12 Effect of variation in irrigation concentration on peak depth and concentration, using maximum, minimum and mean field capacity to determine PEAKM inputs.

As PEAKM is a capacity driven model, soil moisture can only move down through the soil profile when the soils field capacity is exceeded. For this assessment of the model the field capacity is set at a moisture content of 21 percent by volume. Figure 5.13 presents a comparison of soil moisture values at the 120 cm level as predicted by PEAKM and that determined using the matric potential data for this site. This figure clearly displays the importance of the field capacity parameter with its value determining the upper limit of the predicted soil moisture. At the monthly time step, predicted soil moisture values are shown to remain unrealistically at the input level of the field capacity. Statistical summaries of these plots are presented in table 5.11. These figures confirm that PEAK model provides relatively good predictions of mean soil moisture conditions but as displayed in figure 5.13 is unable to reproduce the variations observed in the field.

To examine the PEAK models ability to predict irrigation drainage the soil moisture flux as determined by the model and the three approaches examined in Chapter 4 of this report are compared (figure 5.14). Clearly the capacity approach of this model is more able to simulate the soil moisture flux at DDM03 than the LC and SODICS models. The more rapid macro-pore flow drainage is more easily simulated by the movement of water from one layer to another once field capacity is exceeded. Soil moisture flux values modelled at both the weekly and the monthly time interval are correlated with those determined using the three approaches examined in Chapter 4. As indicated in figure 5.14, the capacity approach of the peak model

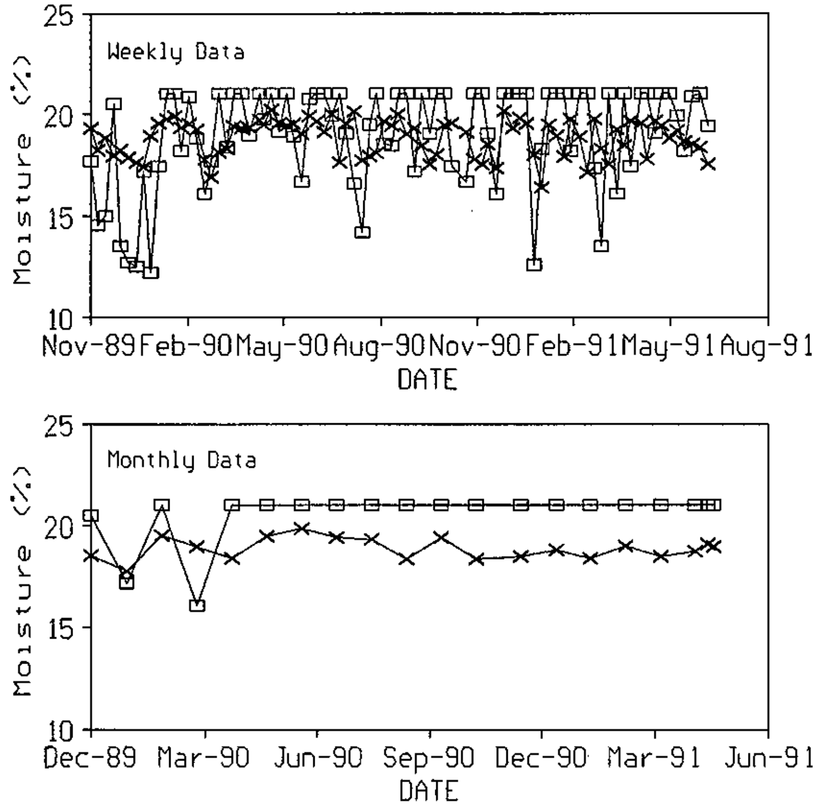


Figure 5.13 PEAKM soil moisture (squares) predictions using weekly and monthly input data compared with observed data (crosses).

Table 5.11 Soil Moisture predicted by PEAKM and measured in the field.

		MAXIMUM	MINIMUM	MEAN	STD. DEV.
Weekly	PEAKM	21.00	12.20	18.98	2.49
	FIELD	20.26	16.40	18.86	0.89
Monthly	PEAKM	21.00	16.10	20.54	1.31
	FIELD	19.86	17.80	18.91	0.51

is more able to simulate the rapid soil moisture movement that occurs as a result of macropore flow. The magnitude of the predicted flux values are very similar to those estimated by the water balance approach for which the best correlations are also shown on table 5.12. However, as discussed in section 4.2, this method of soil moisture flux estimation appears to produce over estimates at the 120 cm depth of the soil profile. The more realistic values of soil moisture flux, estimated by the mass balance approach are greatly exceeded by the PEAK model predictions. Figure 5.15 provides a comparison of the soil moisture flux values determined at monthly intervals and more clearly displays the over predictive nature of the peak model.

Table 5.12 Coefficients of determination for regression equations relating soil moisture flux determined by the PEAK model against three methods of flux measurement examined in Chapter 4.

	MATRIC	WATBAL	MASSBAL
Weekly Data	6.24	50.97	26.53
Monthly Data	8.03	95.06	33.50

where MATRIC = flux measured using differences in matric potentials
WATBAL = flux generated from the water balance approach
MASSBAL = flux generated from the mass balance approach

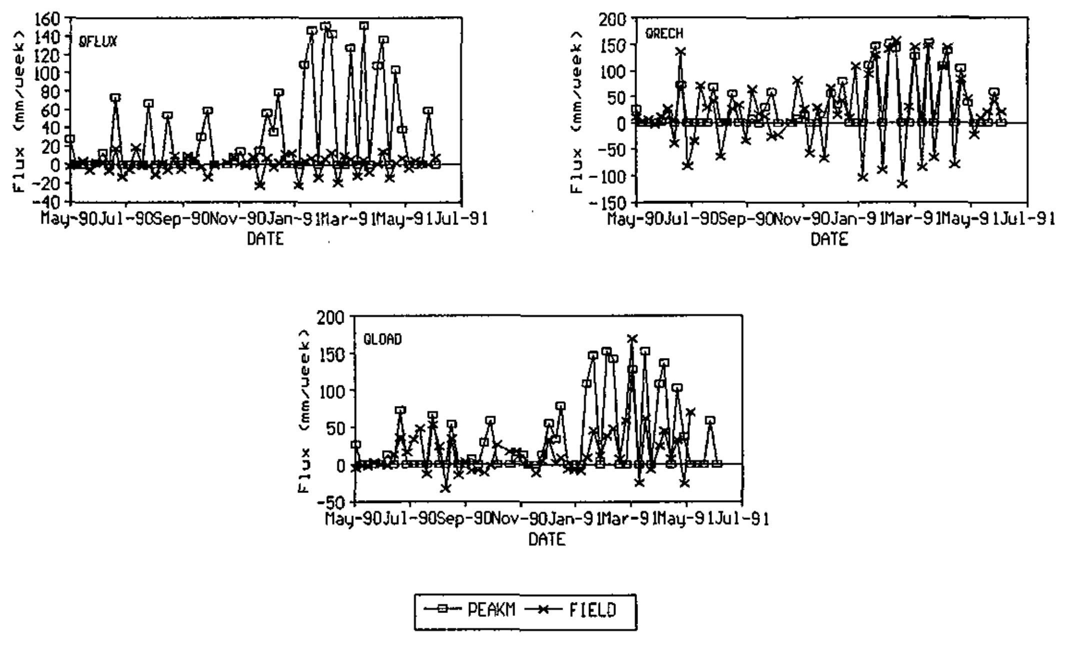


Figure 5.14 Weekly soil moisture flux values predicted using PEAKM compared with that determined by the Darcian (MATRIC), the water balance (WATBAL) and the mass balance (MASSBAL) approaches

Due to the over estimation of soil moisture fluxes by PEAKD, for the purpose of this study where the aim is to demonstrate the need for better hydrosalinity models, PEAKM is not applied to the conditions of the micro-plot DDM03. Although the capacity approach may provide the basis of a useful irrigation management tool, its inability to distinguish between macro-pore and micro-pore flow may prove a major limitation for the development of a research model. Clearly there are two soil moisture processes operating simultaneously within the soil profile. Not only do these process determine the soil water drainage regime, they also provide the major controlling factor in determining solute transport. It may be necessary to model both drainage processes as separate components of the same hydrosalinity model.

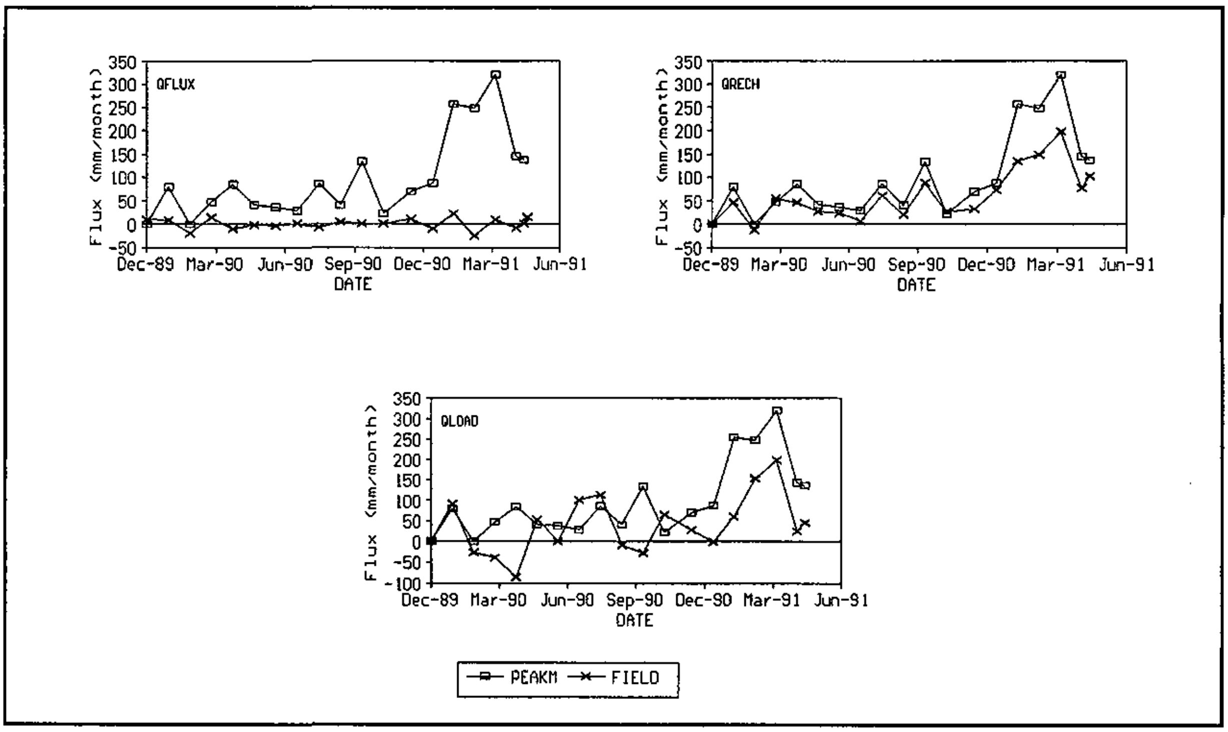


Figure 5.15 Weekly soil moisture flux predicted using PEAKM compared with that determined by the Darcian (MATRIC), water balance (WATBAL) and mass balance (MASSBAL) approaches.

5.4. MODEL LEACHC

LEACHM is a general acronym (Leaching And Chemistry Model) that refers to four simulation models which describe the water regime and transport of solutes in the plant root zone. These models utilize numerical solution schemes to simulate water and chemical movement. They differ in that one model, LEACHN is organized to describe nitrogen transport and transformation, a second, LEACHP, is intended for simulating pesticide displacement and degradation, LEACHC, is formulated to describe transient movement of organic ions (Ca^{2+},Mg^{2+},Na^{+},K^{+},SO_4^{2-},Cl^{-},CO_3^{2-},HCO_3^{-}) and a forth model, LEACHW, only describes the water regime.

The numerical differencing procedures used in this model have been developed from several earlier models (Bresler, 1973; Nimah and Hanks, 1979; Tillotson et al., 1980). The chemical equilibrium and cation exchange subroutines are similar to those of Robbins et al

(1980a and b). The water flow subroutine is based upon that described by Hutson (1983) and uses a numerical solution of the Richards equation as a means of predicting water contents, fluxes and potentials. Improvements to these models include the flexibility of simulating layered or non-layered homogeneous profiles and improved mass balancing. The data is entered by way of a text editor. This aspect of the model could be drastically improved by using the menu and windowing procedures now available for PC's.

The model is very data intensive with over 22 different soil, plant and meteorological variables to which variables must be assigned. If the physical nature or chemical composition of the soil changes with depth the data requirements are even greater as each layer is individually defined. The more important data requirements are:

i. Soil properties and initial conditions with depth.
   a) Depth and thickness of profiles
   b) Water content or matric potential
   c) Hydrological constants for calculating retentivity and hydraulic conductivity.

ii. Soil boundary conditions of:
   a) Irrigation frequency, duration and rate on a daily basis.
   b) Rainfall frequency, duration and rate on a daily basis.
   c) Pan evaporation on a weekly basis.
   d) Groundwater level on a weekly basis if it is in contact with the lower boundary of the profile being modelled.

iii. Upper and lower soil water potentials for extraction by plants.

iv. Crop factors, ie., days from planting to crop maturity, time of harvesting, root distribution and growth.

Due to the very intensive data requirements of this model it must be classified as a research tool and should not be considered for routine irrigation management.

5.4.1 Sensitivity analyses

Due to the intensive data requirements of the LEACHC model it was beyond the scope of this project to carry out a full sensitivity analyses. This study therefore examines the main conclusions from a sensitivity analyses carried out by Moolman (1991). The primary aim of Moolman's sensitivity analyses was to determine which variables play the most important role in determining the quantity and quality of water leaving the root zone. Moolman simulated the water and salt transport in two hypothetical soils at weekly intervals for a period of one year. The two soils differed in that one was saline and the other non-saline. Of the 22 different input requirements for LEACHC Moolman elected to evaluate the following:

i. The a and b coefficients of the Campbell equation (ie. the air entry potential and slope of the soil water characteristic curve).
ii. Saturated hydraulic conductivity (Ksat).

iii. Cation exchange capacity (CEC).

iv. Cation exchange selectivity coefficients, with specific reference to the Ca-Na exchange process (k-Ca/Na).

v. ET/I ratio, which was used as an index of the potential flux of water through the root zone.

The main conclusion from Moolman's study was that one should concentrate on accurately quantifying both evaporation and irrigation rather than attempting elaborate methods to gain accurate values for the chemical and hydrological requirements of the model. However, as the model was found to be most sensitive to the ratio of ET/I it is unfortunate that the importance of these data was not investigated more thoroughly.

The ratio of ET/I was varied by maintaining a constant weekly rate of evaporation and irrigation (both 35 mm/week) and then varying the crop factor from 0.5 to 1.5. Had the application of irrigation been varied over the more realistic range of 0 to 100 mm/week then the model parameters of Ksat and the a and b coefficients of Cambells equation may well have been found to significantly influence model output. One would expect the response of soils with high and low Ksat values to be different with respect to their water and salt load drainage, depending on the amount of irrigation water applied. Under excessive irrigation the soil with a high Ksat would be expected to drain more rapidly and the solute concentration of this drainage water would be less than that from a soil with a low Ksat value. A soil with a low Ksat value should retain more soil water, leading to greater evaporative loss and a greater concentration of the soil water solute load. Not to have examined the importance of the depth and rate of an irrigation application, which are important controlling factors of the soil water and solute movement processes, suggests that the results of this sensitivity analyses must be considered inconclusive.

It is also unfortunate, especially in the context of the current study that such high values of CEC were adopted in the sensitivity analyses. The values determined for the Sundays River area are generally less than 30, the minimum value used in Moolman's sensitivity analyses. It is therefore difficult to interpret the results of this sensitivity analysis in terms of the range of data values found representative of the lower Sundays River valley.

5.4.2. Assessment of LEACHM using field data

To evaluate the LEACHC model data collected for the DDM03 micro-plot in orchard M of the Daisy Dell farm were used to run the model. The predicted soil moisture and soil water drainage from this profile were compared with that determined by the three methods examined in Chapter 4 of this study. Table 5.15 summarises selected input parameters used in this evaluation and the results are presented in figures 5.16 and table 5.17.

At 30cm depth, predicted soil moisture is consistently around 6 percent by volume greater than actual soil moisture (figure 5.16). This disparity is reduced at the 60cm depth, where predicted moisture is approximately 3 percent by volume greater than actual moisture. At

Table 5.13 Selected input parameters for LEACHM.

Input Parameter	Value	
Maximum (actual transpiration/potential transpiration)	3.0	
Molecular diffusion coefficient (mm^2/day)	120	
Dispersivity (mm)	40	
Bottom boundary condition	Free draining	
Particle Size	Clay	Silt
30cm	25%	9%
60cm	27%	13%
90cm	23%	12%
120cm	17%	9%
Bulk Density (Rho) kg/dm^3		
30cm	1.58	
60cm	1.64	
90cm	1.49	
120cm	1.71	
Organic Carbon (%)		
30cm	0.42	
60cm	0.14	
90cm	0.08	
120cm	0.06	
Root Fraction		
30cm	0.6	
60cm	0.3	
90cm	0.1	
120cm	0.0	
Root depth (m)	1.00	
Crop cover	0.6	
Plants per square meter	0.5	
Pan factor	0.76	

90cm the mean moisture values for both predicted and measured data are approximatelyequal, and at 120cm the predicted moisture is 4 percent by volume lower than actual moisture. Overall, the mean predicted soil moisture decreases with depth in the soil, becoming less variable. Below 60cm predicted moisture is constant over time, which is unrealistic and does not replicate the field measurements for the study period.

The inability of LEACHC to accurately predict the absolute soil moisture content is explained with reference to the theoretical basis of the model. Macro-pore flow is not considered by the model as only micro-pore diffusion and dispersion processes comprising the CDE are adopted. The occurrence of macro-pore flow is suggested to be the factor most likely to explain the disparity between absolute soil moisture, both predicted and measured.

Macro-pore flow is most likely to operate within the 30cm layer, where compaction is less severe, and the effects of ploughing are greatest. At this depth the model over predicts the soil moisture content, as moisture draining rapidly through macro-pores is not taken into account. At the 60cm depth, the over prediction of soil moisture by LEACHM is less pronounced. This is because increased bulk density and compaction with depth reduce porosity and hence macro-pore movement is inhibited. With less macro-pore flow the model predictions more closely approximate actual soil moisture, as seen at the 90cm depth where mean predicted and measured moisture are approximately equal.

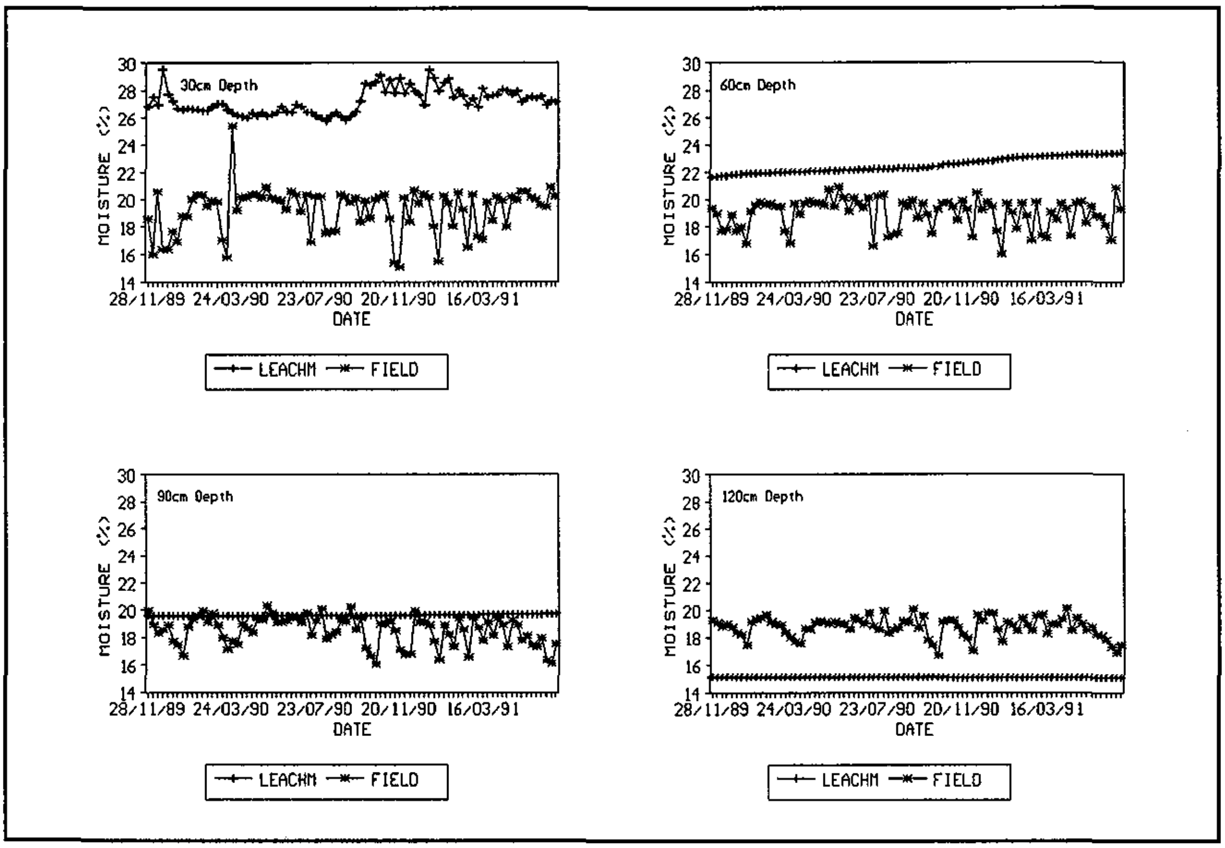


Figure 5.16 Soil moisture predicted using LEACHM compared with field measurements.

LEACHC under-predicts soil moisture at the 120cm depth, by about 4 percent by volume. This may also be by the model inability to account for macro-pore flow. The model therefore tends to accumulate water in the upper layers and therefore under predicts those at greater depth.

Predicted soil moisture flux

LEACHC predictions of soil moisture flux were compared with the values estimated by the three methods discussed in Chapter 4, namely the Darcian method, water balance and Chloride mass balance approaches. Values of leaching flux as predicted by LEACHC are most closely approximated by those determined by the Darcian approach (figure 5.17). This was expected as the water flow routines of LEACHC are numerical solutions of the Richards equation which only considers micro-pore flow. Figure 5.17 provides a comparison of the leaching flux as determined by LEACHC and that determined using the Darcian method. Clearly when one considers the inability of the Darcian approach to accurately estimate soil moisture flux at DDM03 as discussed in Chapter 4, the use of LEACHC is most inappropriate for modelling soil moisture and solute transport in these soils. To further assess the utility of this model for use in soils of the Sundays River valley is considered unproductive.

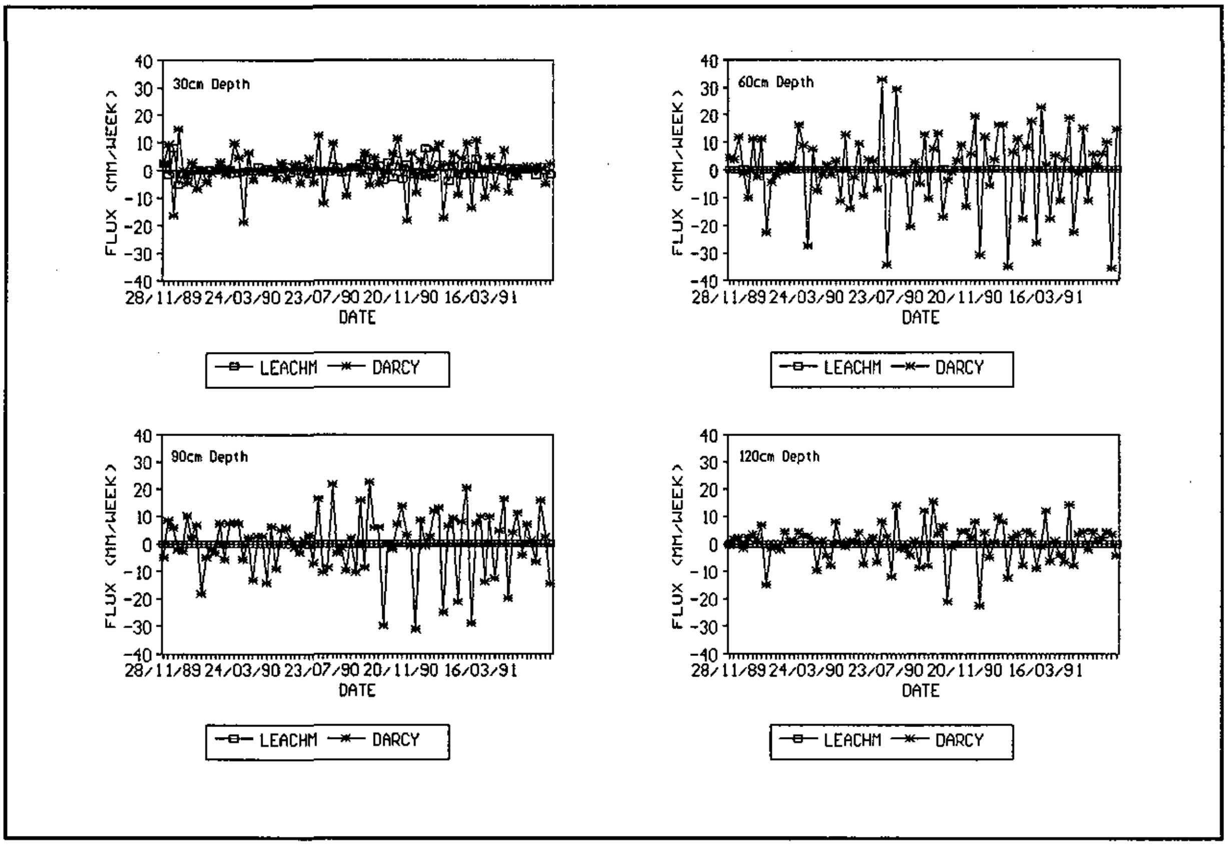


Figure 5.17 Soil moisture flux predicted using LEACHM and determined using Darcy's Law.

5.5. CONCLUSIONS AND RECOMMENDATIONS

Within Chapter 4 of this report the importance of macro-pore flow as a component of irrigation drainage was clearly demonstrated. It is therefore unreasonable to expect hydrosalinity models that do not consider this phenomena to be applicable to the relatively well draining alluvial soils of the lower Sundays River valley. This chapter set out to evaluate a number of hydrosalinity models and has demonstrated that they are inappropriate for application in the study area rather than provided recommendations on their use. This illustrates the need to identify the major factors controlling the processes of soil moisture drainage and solute movement before selecting and applying hydrosalinity models. This also highlights the need for further research to develop hydrosalinity models that do account for rapid macro-pore flow which is often the major component of irrigation drainage, especially for alluvial soils which are commonly used for intensive irrigation farming.

Three hydrosalinity management models of differing theoretical complexity and having different levels of input data requirements, and one research hydrosalinity model were simply assessed to determine their appropriateness for predicting irrigation drainage from soil of the lower Sundays River valley. The models were all assessed using data collected for the micro-plot site of DDM03 to determine leaching flux.

There are three common reasons for the failure of hydrosalinity models:

i. Incorrect computer coding

ii. Poor understanding of the processes to be modelled

iii. The selection of an inappropriate model.

Of the models assessed in this section of the study all were found to compile and run satisfactorily although Moolman (1991) does express some frustration due to run time errors when attempting to apply the LEACHC model to relatively neutral soils. The model seemed to exhaust the supply of CO_3 and develop run time errors related to division by zero. This problem was not experienced with the alkaline soils of the Sundays River study area. Therefore, in the context of the current study, the first of the three major problems related to the application of hydrosalinity models as listed above is not considered relevant.

However, the second and third of the three problems listed above deserve further discussion. When one considers that a large proportion of the intensive irrigation farming within southern Africa is on well draining alluvial soils it is difficult to understand the persistence of salinity research to focus on further applications of the Richards equation rather than examine and develop modelling techniques that consider macro-pore flow. There is little excuse for not recognising the importance of macro-pore flow in any particular situation when this can be easily assessed using simple water budget or mass balance approaches. The major problem would therefore seem to stem from the application unappropriate of models. This may be attributed to the unavailability of appropriate hydrosalinity models that adequately deal with macro-pore flow, and the acceptance of models developed for less well drained soils where micro-pore flow dominates. The significant efficiencies in use of irrigation water that may my be gained through improving current knowledge on irrigation drainage and the development of new irrigation management strategies should provide adequate motivation for research managers within local to develop hydrosalinity models more appropriate to South African conditions.

6. GROUND WATER WITHIN THE LOWER COERNEY VALLEY

The lower Coerney valley comprises a system of alluvial terraces overlying the Kirkwood and Sundays River formations. These mudstone formations are of marine and fluvial origin and overlie the Table Mountain sandstones. A more detailed description of this geology is presented in Section 2.2. Within the alluvium there is a localised ground water system which is highly saline and of limited use for agriculture. A number of boreholes were drilled for this project by the Dept. of Water Affairs and Forestry to provide information on both the level and salinity of the ground water within this small alluvial aquifer. Data from these boreholes have proved most useful in determining the impact of irrigation on both the hydrological system and in particular on the ground water of the lower Coerney valley. Before examining the ground water within the lower Coerney valley a brief description of the alluvial deposit is provided. Variations in the level of ground water and its salinity are then examined and finally the ground water system of the valley is described.

6.1 GROUNDWATER DATA COLLECTION

To provide information on the level and salinity of ground water within the lower Coerney Valley the Department of Water Affairs and Forestry drilled a number of boreholes through the alluvium and just penetrating the underlying cretaceous mudstones (figure 6.1). The holes were screened with perforated PVC pipes and backfilled with small stone chip. Initially the level of ground water was monitored continuously at seven sites using electronic data loggers. A further five sites were monitored manually. However, after the first two years of the project the MC System's data loggers used for the continuous recording of water levels were found to be unsatisfactory and subsequently all ground water monitoring was carried out manually using an electric probe. Currently 20 boreholes are monitored on a monthly basis. Their rest water levels are measured, surface water samples are collected and periodically the electrical conductivity profile of the water within each borehole was determined using an electrical conductivity meter with a probe fixed to an extended cable. Information from the drilling of these holes was also used to gain information on the alluvial deposit of the lower Coerney valley.

6.2 THE ALLUVIAL DEPOSIT OF THE LOWER CORENEY RIVER VALLEY

The geology of the lower Coerney River valley comprises a series of alluvial terraces over lying the cretaceous mudstones of the Kirkwood and Sundays River formations. An examination of the geological logs compiled for the boreholes drilled into the alluvial deposit of the lower Coerney valley show that the aquifer is comprised of mainly sandy material mixed with varying quantities of either loam or clay. In some instances well rounded quartzite pebbles are associated with sandy layers. These pebbles are poorly sorted and range in size from 10 to 100 mm in diameter. Numerous clay lenses are randomly distributed throughout the vertical profile of each borehole. These clay lenses are made up of varying percentages of silt, clay and sandy material. An attempt was made to horizontally correlate the different soil layers identified within each borehole, but this proved unsuccessful and no common stratigraphy could be identified with the density of borehole information available.

An overall evaluation of the borehole profiles indicate no uniform depositional pattern across the aquifer. This can be explained by the continued lateral migration of the Coerney River across the valley accompanied by periods of flooding onto the flood plain. The lateral migration of the meandering channel has led to the deposition of point bars and their associated course grained deposits. The clay and silt layers may have resulted from over bank flooding and subsequent deposition of finer material. This continual process of deposition and erosion has resulted in the formation of these isolated lenses.

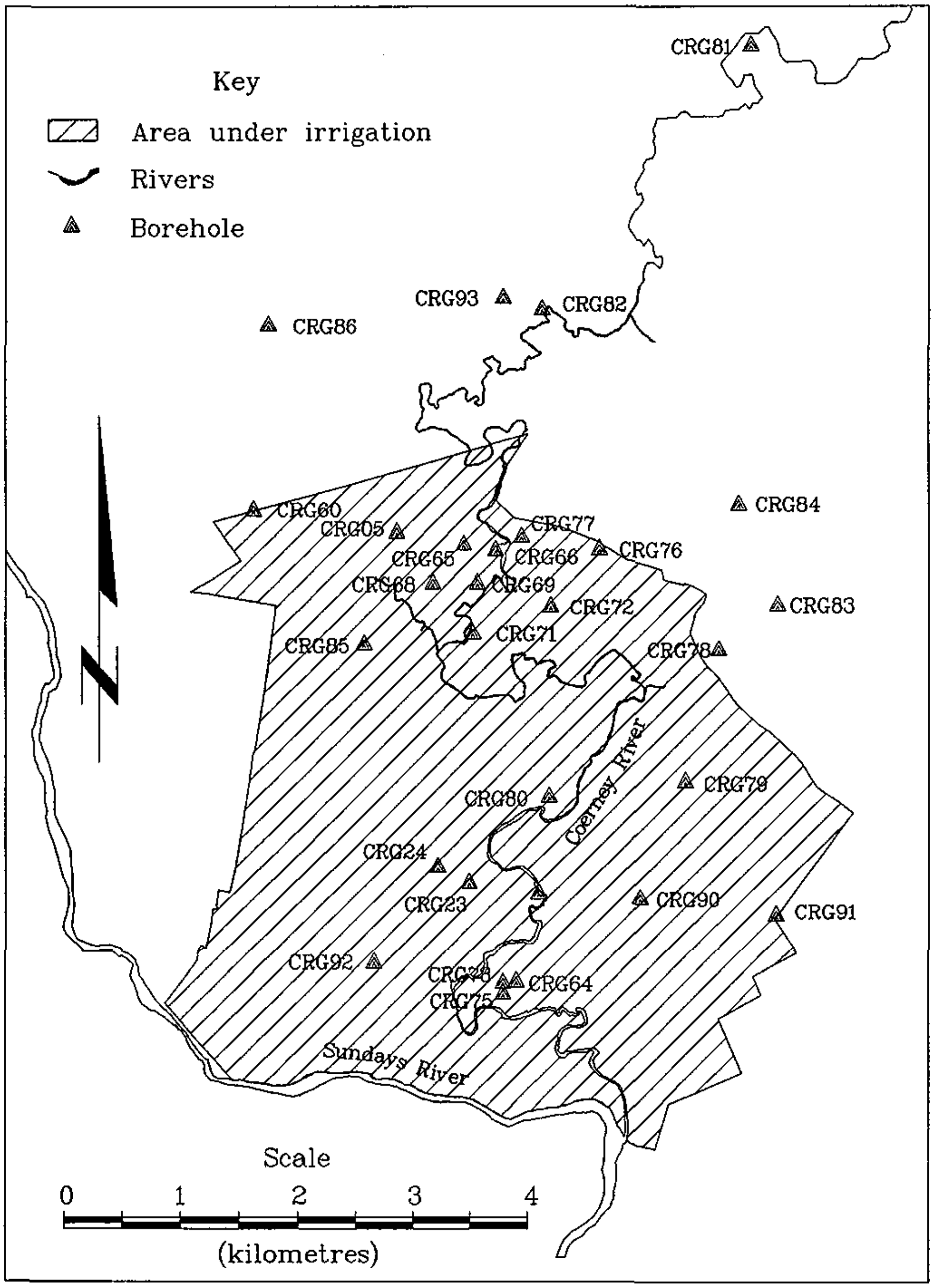


Figure 6.1 Location of groundwater sampling sites within the lower Coerney Valley

The heterogeneity of the alluvial deposit has important implications as regard the hydraulic properties of the aquifer. Course-grained unconsolidated sand deposits form the principle water storage medium with the associated pebble beds enhancing the hydraulic conductivity of this alluvial material. However, these zones of high hydraulic transmission are often interlayered with lenses of silt and clay which reduces both the overall water yielding capacity of the aquifer and its transmissivity.

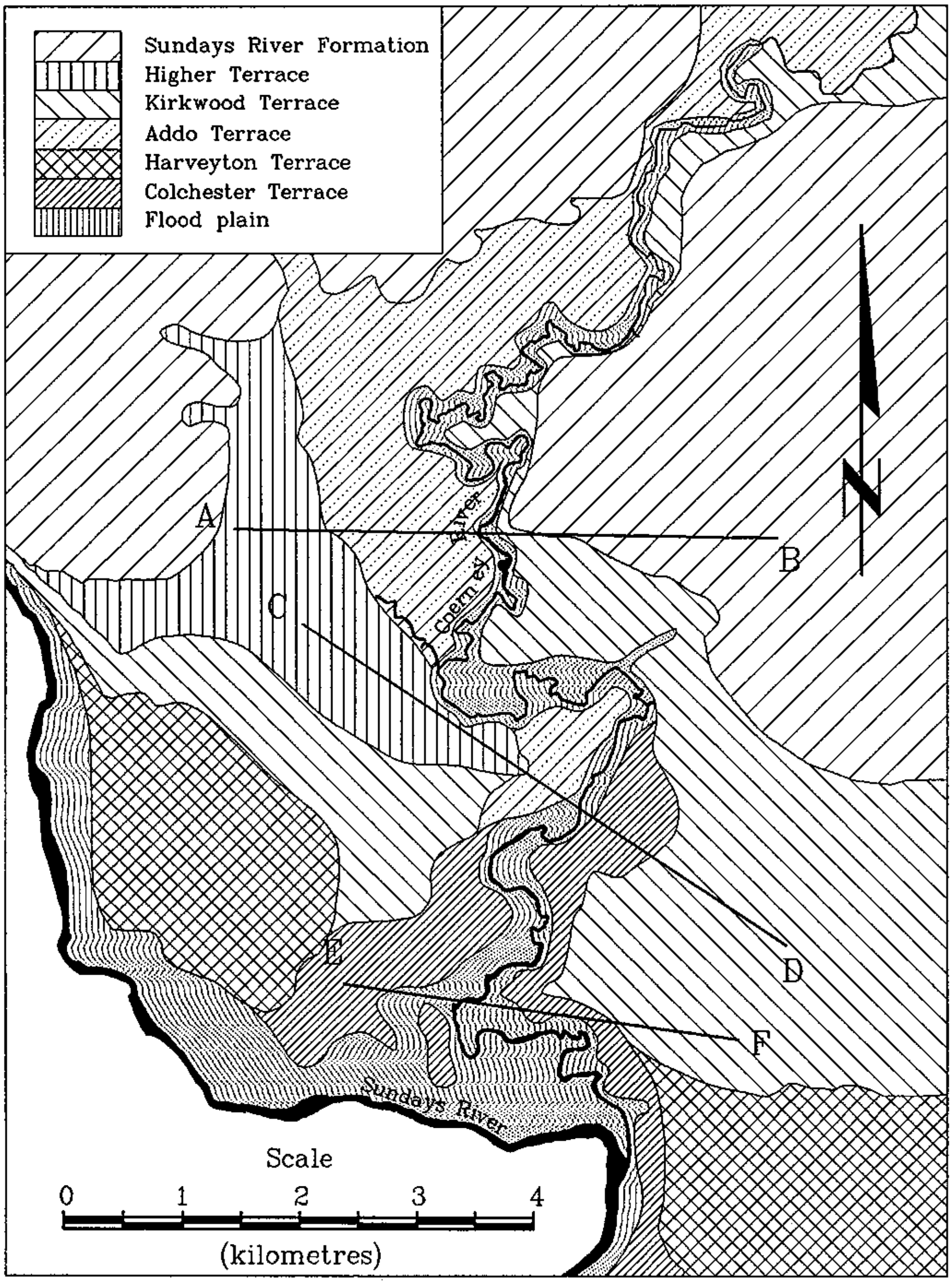


Figure 6.2 Location of cross-sections through alluvial deposit of lower Coerney valley as draw on figure 6.3

The depth of contact between the Cretaceous bedrock and the overlying alluvium has been estimated from the borehole logs and is used to determining the thickness of the aquifer. The aquifer thickness ranges from less than 15 metres in the upstream reaches of the aquifer to nearly 30 metres further downstream. From the borehole information, relief of the upper surface of the cretaceous formations appears relatively smooth with no major sinks. The buried bedrock channel appears to have a trough-like shape with the alluvium gradually thinning towards the margins of the aquifer (figures 6.2 and 6.3).

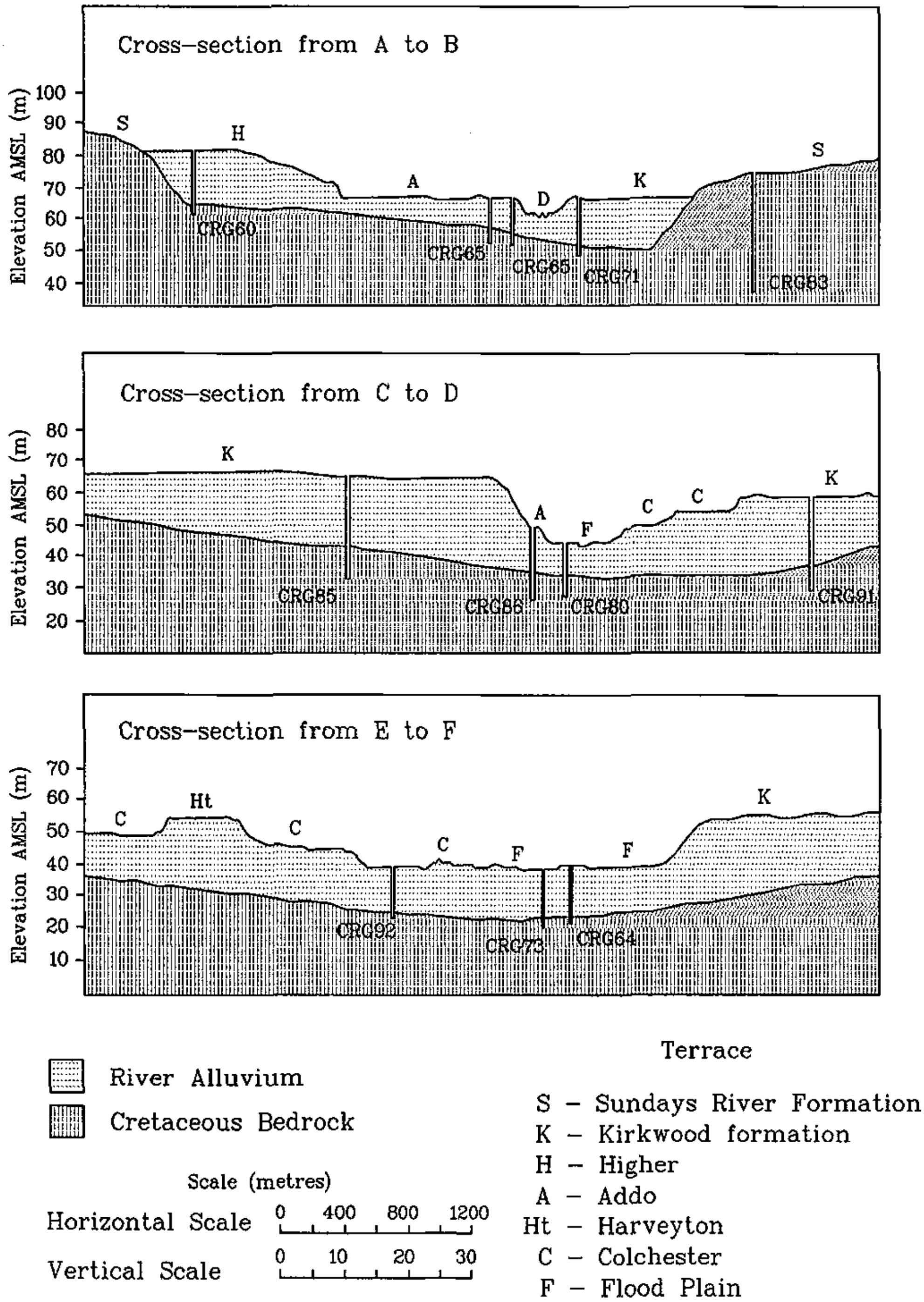


Figure 6.3 Cross-section profiles through the alluvial deposit of the lower Coerney valley (see figure 6.2)

6.3 SPATIAL VARIATIONS IN THE GROUNDWATER LEVEL

To examine spatial variations in ground water level of the lower Coerney valley a contour map of water table elevations was constructed. This map was constructed using a surface interpolation and mapping package. Ground water levels as recorded in May 1990 were used to construct an isoline map of the ground water elevation above mean sea level (figure 6.4). From borehole CRG81 (figure 6.1) the ground water level is shown to steadily decrease in a down valley direction towards CRG82. This

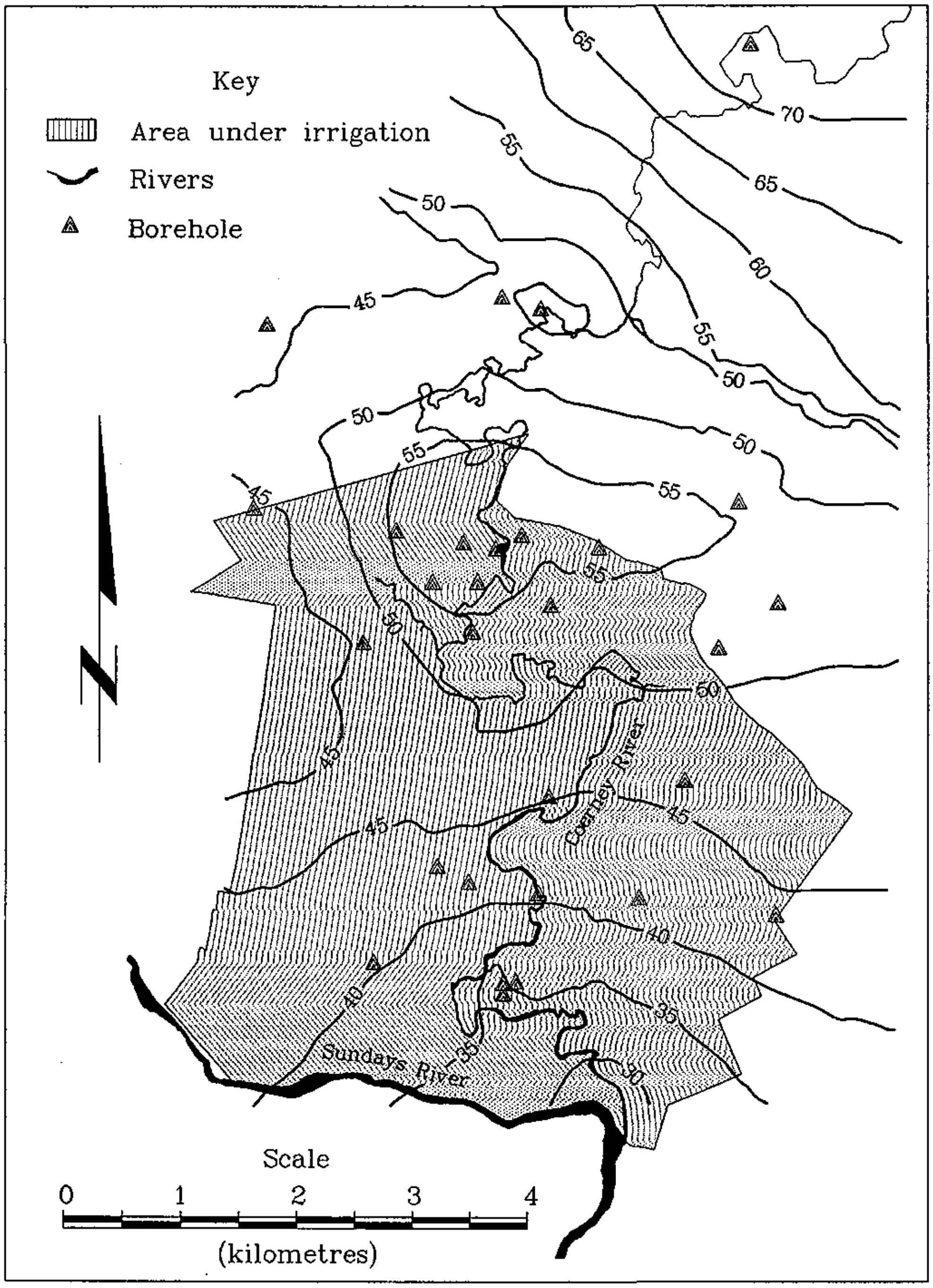


Figure 6.4 Groundwater levels of the lower Coerney valley displayed as 10 metre isolines above mean sea level.

fall in elevation is in sympathy with the falling topography and is unlikely to represent major changes in the hydraulic characteristics of the alluvium. Between these sites the alluvial deposit is generally narrow and relatively shallow. However, as one moves further down valley the water table begins to rise in elevation as one approaches the up valley extent of the area currently irrigated within the lower Coerney valley. At CRG68, which is located near the centre of a large field, irrigated with a centre pivot overhead spray system, the water level rises to approximately 68 m above mean sea level. This is within 1.2 metres of the land surface and causes major problems in terms of crop production. Down valley of borehole CRG78 the level of the ground water decreases with the fall in valley elevation to less than 30 m near the confluence of the Coerney and Sundays Rivers. It is clear that within the area of intensive irrigation the ground water level has risen significantly. It should also be noted that the water table on the right bank of the Coerney River is significantly higher than that on the left bank where irrigation farming is less intensively developed.

6.4 TEMPORAL VARIATIONS IN THE GROUNDWATER LEVEL

To determine temporal variations in the level of ground water within the lower Coerney valley time series plots of the ground water level are examined (figures 6.5 to 6.8). These plots are organised with borehole records grouped according to elevation with the same vertical scale being adopted for each figure. An initial inspection of these time series plots reveals three types of ground water regime, those with falling, stable and rising levels.

Borehole CRG23, CRG64, CRG73, and CRG92 provide information representative of the ground water level in the lower Coerney valley on either the flood plain of the Sundays River or on the Colchester Terrace. This area has one of the longer histories of irrigation within the Coerney valley. Other than that for CRG64, the borehole records for this area show the ground water level to be relatively stable. The record for CRG64 is unique in that it shows a significant fall in the water table. This is attributed to a localised change in land management from intensive flood irrigated lucerne and maize to unirrigated pasture.

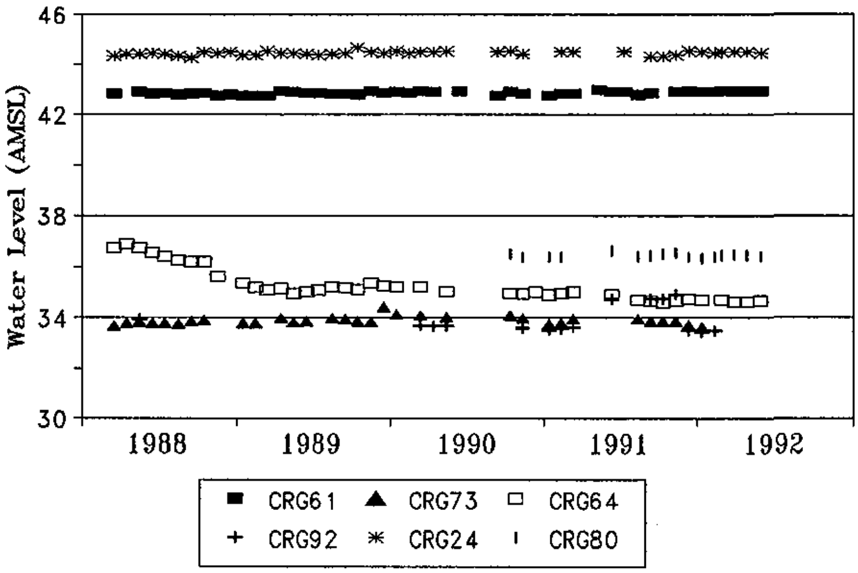


Figure 6.5 Groundwater level in boreholes CRG05, CRG68, CRG69 and CRG71.

The boreholes on the left bank of the Coerney River and within the area currently developed for irrigation are mainly drilled through the Kirkwood Terrace. This area, especially that up valley of borehole CRG90 has under gone major irrigation development during the period of this study. The impact of this development is clearly shown by the steady rise in the water table of this area. This trend is especially evident in the ground water record for boreholes CRG76 and CRG77. However the trend is less evident for CRG71 which is located on the edge of the Kirkwood Terrace adjacent to the Coerney River

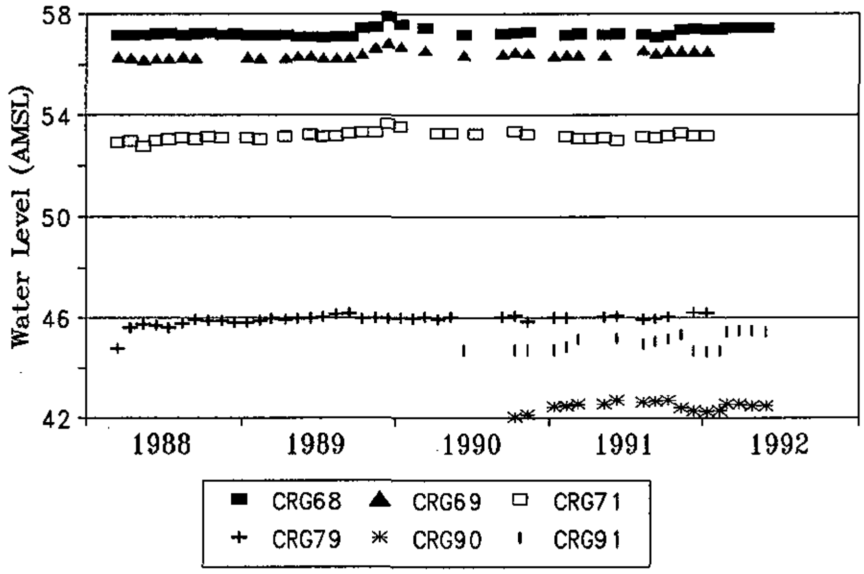


Figure 6.6 Groundwater levels in boreholes CRG68, CRG69, CRG71, CRG79, CRG90 and CRG91.

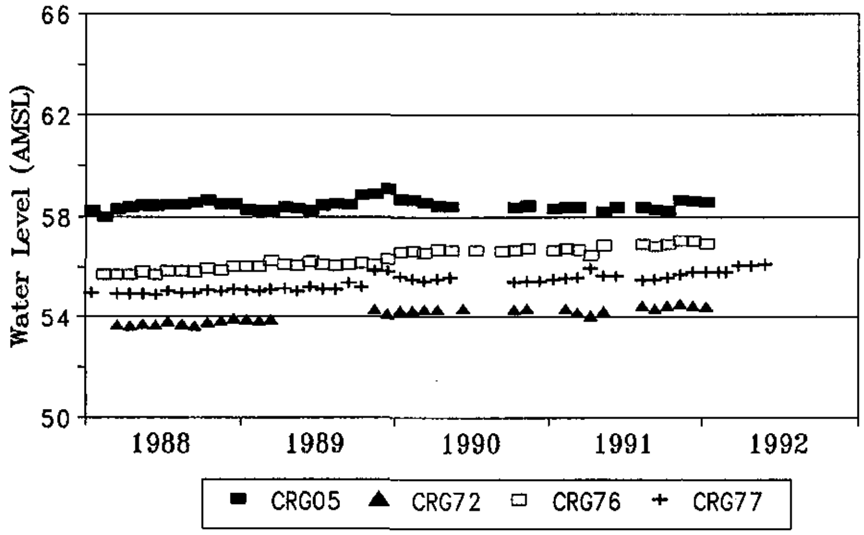


Figure 6.7 Groundwater levels in CRG05, CRG72, CRG76 and CRG77

Another definable area of the valley that has undergone major irrigation development is located on the right bank of the Coerney River up valley of borehole CRG85. Borehole CRG68 is located near the centre of a large field initially developed for cash cropping of potatoes and maize during 1986. This area was heavily irrigated with an overhead centre pivot spray irrigation system. Subsequently this area has been planted with citrus and is now irrigated with a micro-jet system. Two further centre pivot irrigation systems have been constructed on the lands between borehole CRG85 and CRG60. These lands are currently used for potatoes and maize production but may well be converted to citrus in the near future. A large farm dam has also been constructed on the edge of the higher terrace to the west of CRG05. The ground water level within this entire area, other than that at CRG05, has displayed a steady rise during the period of the current study. Borehole CRG05 is located at the edge of a well established micro-jet irrigated orchard. Although the water level of CRG05 is higher than that of the nearby CRG68 the level has remained fairly constant during the current study. The water level at CRG60 is even higher. This is not only attributable to the nearby centre pivot irrigation system in this area but also to the higher elevation of the alluvium base nearer the margins of the valley.

There are three boreholes located up valley of the area currently developed for irrigation farming within the valley. Borehole CRG81, the further-most up valley ground water monitoring site used in this study clearly displays a falling level of ground water. This borehole is located sufficiently upstream not to have been effected by the irrigation development within the valley. The falling level of ground water can only be attributed to the drought conditions and resulting lack of recharge during the period of this record. It is unfortunate that this borehole was not drilled prior to the major rainfall and flood event of November 1989. This trend of a falling ground water is also evident, but to a lesser extent, at borehole CRG82. This borehole is located upstream of the area under irrigation but at a point where the generated isoline map of water levels within the valley begins to rise. This suggests that the ground water level at CRG82 may be influenced by water from the irrigated area and explains why its level has not fallen in a similar way to the level at CRG81.

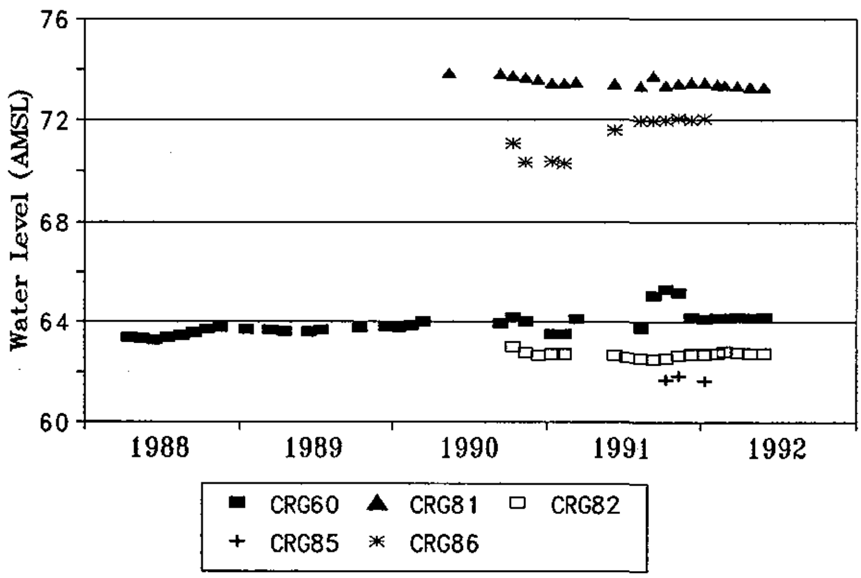


Figure 6.8 Groundwater levels CRG60, CRG81, CRG82, CRG85, CRG86

CRG86 is also up valley of the area under irrigation, but on a tributary arm of the alluvium which drains the valley containing the new Scheeper Vlakte dam. It is of interest to note that the level of ground water recorded at this borehole underwent a rapid rise following the initial filling of the Scheepers Vlakte dam. This suggests that seepage from this dam may provide a major new recharge source to the alluvial aquifer of the lower Coerney valley.

Due to changing patterns of irrigation and farm management practices, as well as the spatial interaction of ground waters of different salinity characteristics, it is difficult to fully explain most short term temporal variations in ground water variations. Even when examining water level data collected on a weekly basis seasonal variations of water level are not readily apparent. However for most sites the water level record clearly displays a response to the major rainfall and flood event of November 1989. It is of interest to note that the rise in water table from this recharge event was not sustained and that within a very short time water levels had fallen to prevent levels. This may indicate that the level of the ground water in of the Coerney valley is determined by an equilibrium between the rate of recharge from irrigation and drainage from the alluvium. There is no indication that the observed rising water table is due to drainage problems leading to widespread water logging problems.

6.5 SPATIAL VARIATIONS IN GROUNDWATER SALINITY

To examine spatial variations in ground water salinity of the of the lower Coerney valley an isoline map of salinity levels was constructed. This map was constructed using a surface interpolation and mapping package. Salinity levels as recorded in May 1990 were used as data to construct this isoline map of ground water salinity expressed in terms of electrical conductivity ($mS.m^{-1}$). Before examining the resultant isoline map it is important to recognise that the shape of the salinity profile between two data points is determined by the adopted method of interpolation. Especially within areas of sparse data, major jumps in salinity may be smoothed giving a false impression that two very different water bodies are mixed over a wider area than actually occurs in nature. However, this form of analyses and presentation does provide information on more general trends and can indicate steep salinity gradients where data of a sufficient spatial resolution are available.

From borehole CRG81 (figure 6.1) the ground water salinity is shown to steadily decrease in a down valley direction towards CRG82 (figure 6.9). This salinity gradient as portrayed on Figure 6.9 is most probably unrealistic. As displayed on Figure 6.4 and discussed in section 6.3 the ground water level at CRG82 is influenced by recharge from irrigation water applied further down valley. A more realistic salinity profile would be for the high values of salinity to extend down valley from CRG81 to a some point not far up valley from CRG82, the up valley extremity of the area influenced by water recharged into the aquifer from the irrigation development. At this point the salinity of the ground water would decrease more rapidly than suggested by the isoline plot. However, with currently available data it is not possible to more clearly define the upstream extent of the influence of the irrigation drainage. Down valley from CRG82 the salinity continues to decrease to levels of between 400 to 600 $mS.m^{-1}$. However the map clearly defines an area of high salinity centred near borehole CRG71 with electrical conductivity values of approximately 2000 $mS.m^{-1}$. This salinity value is similar to that measured for CRG81, located well upstream of any influence of irrigation farming, and also of the water continually seeping into the

Coerney River from its left bank between CRQ08 and CRQ07. At CRQ20 this seepage water has been collected using a perforated pipe and regularly sampled. It is interesting to note that CRQ20 is immediately down valley from CRG71. Down valley from this area of very saline ground water the salinity decreases to levels of between 400 and 600 mS.m^{-1} on the flood plain of the valley near its confluence with the Sundays River. Clearly the salinity of the water recharging the ground water system of the lower Coerney valley is less saline that the natural ground water of the area. Information suggesting a reason for the high salinity in the

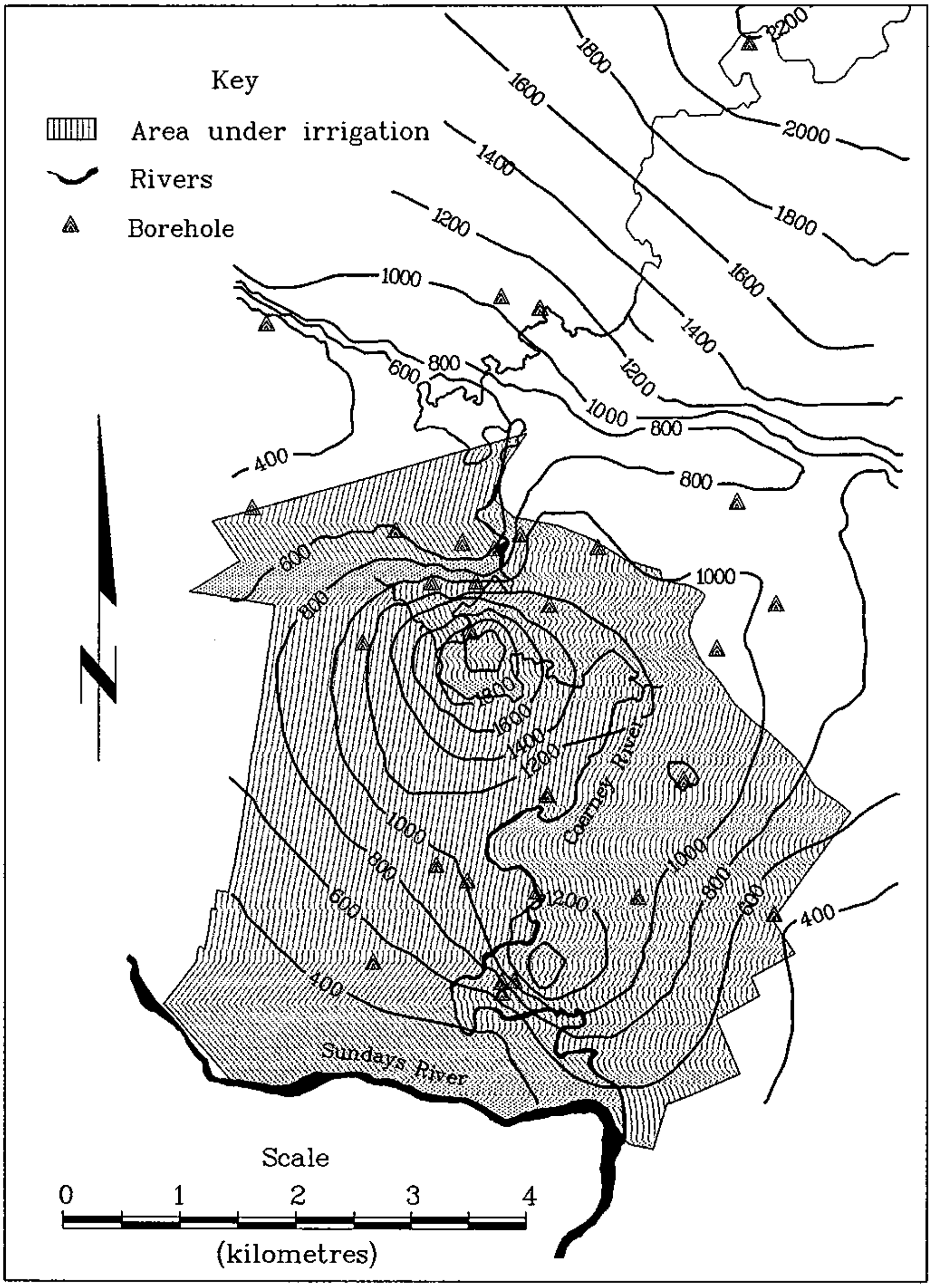


Figure 6.9 Spatial variations of salinity of the lower Coerney valley

close proximity of CRG71 was fortuitously gained during the drilling of a further series of boreholes in an orchard of the farm Daisy Dell. Daisy Dell farm is located on the right bank of the river meander between river monitoring sites CRQ08 and CRQ07. A borehole drilled through the alluvium and penetrating the underlying cretaceous mudstone below struck a weak artesian flow with a salinity of approximately 1500 $mS.m^{-1}$. This water rose to an elevation of nearly 2.5 metres above the ground surface. Unfortunately it was not possible to fit a piezometer or sampling tube to this hole and so the salinity and piezometric head of water were not monitored. A number of other holes were drilled in close proximity to this hole but without making contact with artesian water. It should also be noted that near CRG82 a borehole of approximately 300 m depth and penetrating the Table Mountain sandstones has artesian flow. Unfortunately this hole is in a state of disrepair with the steel casing badly corroded in the highly saline water of the alluvial aquifer. It therefore seems that the cretaceous mudstones of the Kirkwood and Sundays River Formations act as an aquiclude with an artesian aquifer within the Table Mountain sandstones below and a localised aquifer within the alluvial deposit of the Coerney valley above. The water within the Table Mountain sandstone is not saline but as it move up through weaknesses within the marine laid mudstones of the aquiclude, the water becomes highly saline. This water recharges the alluvial aquifer, especially at points of weakness, to create areas of highly saline ground water such as in the area around borehole CRG71.

To further explore the spatial variations of salinity within the alluvial aquifer of the lower Coerney valley the electrical conductivity through vertical profiles of the aquifer were monitored at irregular time intervals. An electrical conductivity probe was lowered down the boreholes on a length of co-axial cable marked with half metre graduations. To improve the resolution of the depth measurements of this monitoring technique distance between the half metre graduations were determined using a metre tape. It should also be noted that inaccuracies could arise due to the increased resistance of the long length of co-axial cable used to connect the electrical conductivity probe to its meter. This required some modifications to the electronics of the conductivity meter. Also due to the extraordinary range of salinity measured down each profile, problems were experienced in accurately calibrating the electrical conductivity meter. Therefore, although the electrical conductivity measurements were stable and consistent between profiles, the absolute values of the data are not directly comparable to those samples analysed in the laboratory. Although collecting this information was problematic the recorded salinity profiles provide a good insight into the three dimensional dynamics of the hydrosalinity processes within this aquifer.

Figure 6.10 presents a representative selection of the salinity profiles through the alluvial aquifer of the lower Coerney valley. An initial inspection suggests three major profile curve types. Firstly, there are those curves of relatively high salinity through out the profile with only minor salinity gradients near the aquifers upper surface. This type of profile was recorded at boreholes CRG71 and CRG81. The second curve type displays a major salinity gradient near the upper surface of the aquifer. A third curve type is that of CRG76 which is characterised by a thick layer of surface water of relatively uniform salinity. This layer of water sits above a major salinity gradient to a more saline water below.

Borehole CRG81 is well up valley from the area currently used for irrigation and at an elevation where it cannot be influenced by irrigation drainage. Borehole CRG71 is at the centre of a localized area of high salinity which has earlier been attributed to the recharge of the alluvial aquifer by saline water rising up through the underlying cretaceous mudstones. The water at both these sites is very similar and does not appear to be significantly influenced by mixing with irrigation drainage.

The second group of profiles comprise those boreholes CRG69, CRG82 and CRG90 which display major salinity gradients near the upper surface of the aquifer. The salinity profile defined for site CRG82 suggests an inflow and mixing of less saline ground water near the upper surface. An explanation for the source of the upper less saline water may be the up valley movement of irrigation drainage water. Being less dense than the significantly more saline natural ground water of the valley, the irrigation drainage water will tend to float as a pool of less saline water near the surface of the aquifer. Figure 6.4 clearly indicates a significant rise in the water table near the up valley extremity of the irrigated area. This rise may comprise a build up of less saline water near the surface which will expand up valley of the irrigated area as a thinning wedge of less saline water. The thinness of this wedge as measured at CRG82 would suggest that this site is near the up valley extremity of the influence of irrigation drainage.

The profile at CRG69 is more complex with two distinct salinity gradients. The first gradient is near the upper surface of the aquifer, suggesting an interface between locally generated irrigation drainage water and that comprising a composite of mixed irrigation drainage and the natural ground water. This composite water which has a salinity ranging from 1000 to 1200 $mS.m^{-1}$ is representative of the main body of water within the aquifer comprising varying concentrations of irrigation drainage and highly saline natural ground water. At CRG69 this water lies above a layer of the very saline (2000 $mS.m^{-1}$) natural ground water. This water is very similar to that of CRG71, the natural ground water which moves up through the saline mudstones of the cretaceous Sundays River and Kirkwood formations.

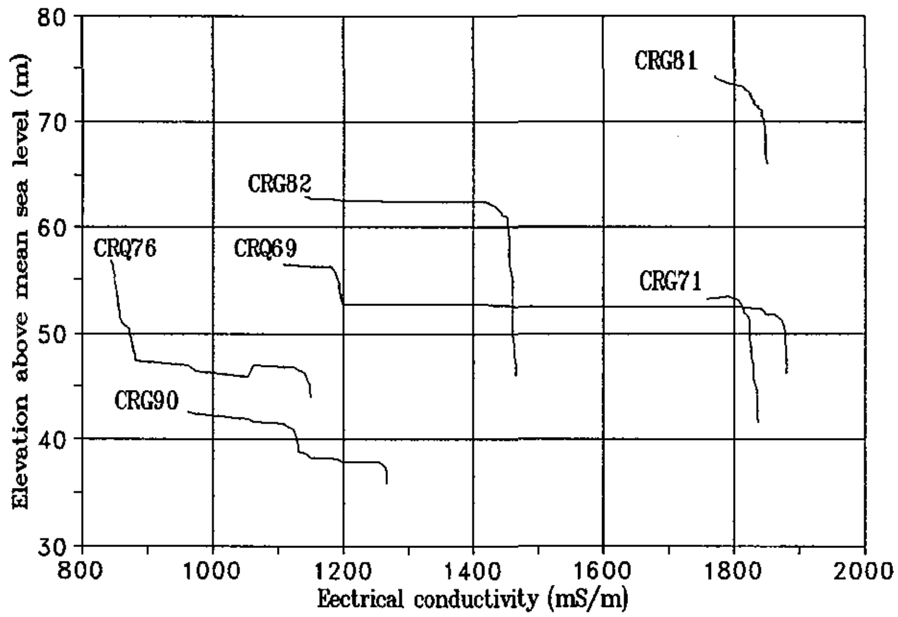


Figure 6.10 Representative profiles of salinity through water columns of the alluvial aquifer in the lower Coerney valley.

CRG69 is located near the right bank of the Coerney River approximately 100 metres upstream from CRG71 which is located on the left bank of the river. Although as yet unsubstantiated it would appear that the upper level of the saline water at borehole CRG69 is controlled by the level of the river which separates this area from that near borehole CRG71, the source of the highly saline ground water.

CRG90 which is located on the Kirkwood terrace at the lower end of the valley also displays a salinity gradient near the aquifer surface. This suggests a mixing of both the irrigation drainage with more saline water deeper in the aquifer. However as this deeper ground water is significantly less saline than that found near CRG71 it is not thought to be emerging through the underlying mudstones at this point.

The salinity profile at CRG76 is unique as it displays no real salinity gradient near the surface. This borehole is located between lands that are currently irrigated and the main irrigation canal. Currently there is no clear explanation for the salinity profile at this site. As the upper layer is relatively thick and significantly more saline than the irrigation water it is unlikely that the profile indicates any significant leakage from the irrigation canal.

By monitoring and then comparing the salinity profiles through the alluvial aquifer the complexity of the system is clearly demonstrated. It also offers a warning to geohydrologist not to assume that ground water samples collected near the upper surface of the aquifer are representative of the entire vertical water profile. In fact, as clearly demonstrated at borehole CRG69, the salinity of the upper water may be only a fraction of that near the base of the aquifer. Unfortunately no profiles were determined subsequent to January 1991. Although the connection between the electrical conductivity probe and co-axial cable were well sealed it prove impossible to completely prevent water from penetrating into the cable and probe connections. Although this equipment had been replaced several time prior to the end of 1990 to continue this practice became too expensive. Therefore with the failure of this equipment during January 1991 no further salinity profile measurements were determined.

6.6 TEMPORAL VARIATIONS IN GROUNDWATER SALINITY

To examine temporal variations in ground water salinity, time series plots of both electrical conductivity and water level as recorded for a number of representative boreholes are examined. Clearly from the examination of spatial variation in both ground water levels and salinity there are a number of different processes operating which are dominant in different parts of the study area. The ground water monitoring sites selected for the following analyses are those for which salinity profiles were examined in the previous section of this report. Unfortunately, as was clearly shown by examining the salinity depth profiles, values of salinity measured at the upper surface of the aquifer may provide misleading information in terms of the aquifer as a whole. Bearing this in mind, it is often difficult to fully understand the temporal variations of salinity at the upper surface of the aquifer.

The time series plot for CRG81, the borehole located well up valley from the influence of irrigation drainage displays a steady decrease in water level and a significant rise in salinity (figure 6.11). The salinity rises to a peak of nearly 2800 $mS.m^{-1}$ during the first half of 1991 before falling to levels of between 2400 and 2600 mS.m^{-1}. With information currently available it is difficult to provide a clear explanation for these variations. However, it seems reasonable to attribute the general rise in salinity to a lack of surface recharge during the current prolonged period of drought. This would cause the observed drop in the level of the water table and a further concentration the soluble salts.

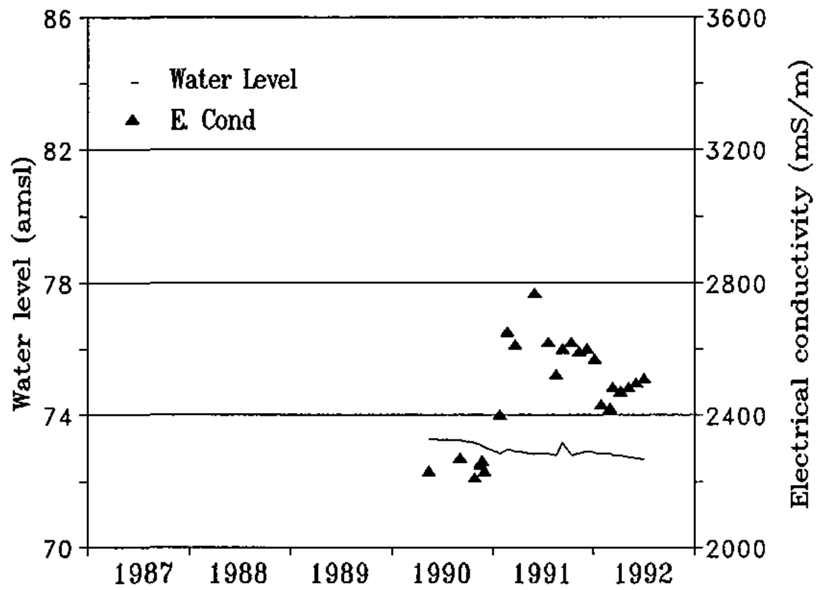


Figure 6.11 Water level and electrical conductivity at borehole CRG81.

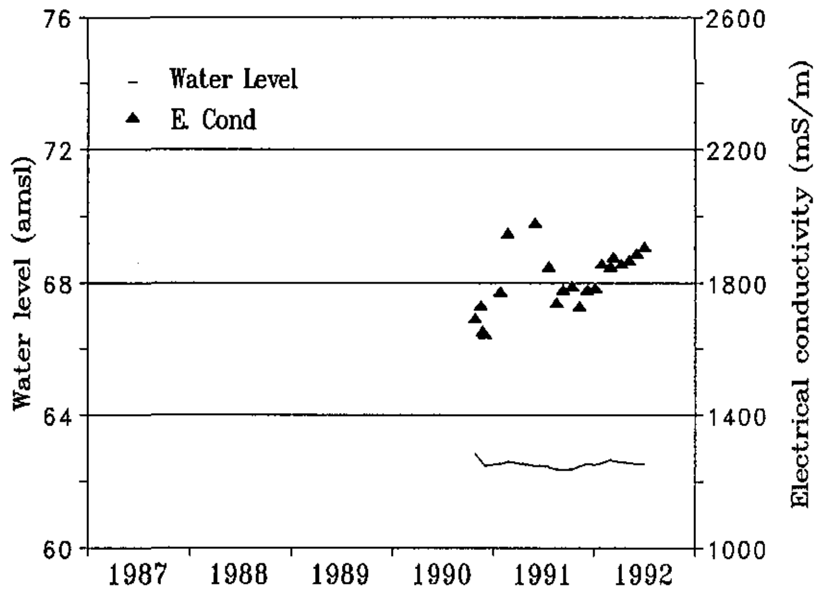


Figure 6.12 Water level and electrical conductivity at borehole CRG82.

It is interesting to note that variations in the salinity of the upper surface of the aquifer at CRG82 follow a similar trend to that at CRG81 (figure 6.2). The reasons for this pattern of variation are therefore thought to be similar to those of CRG81.

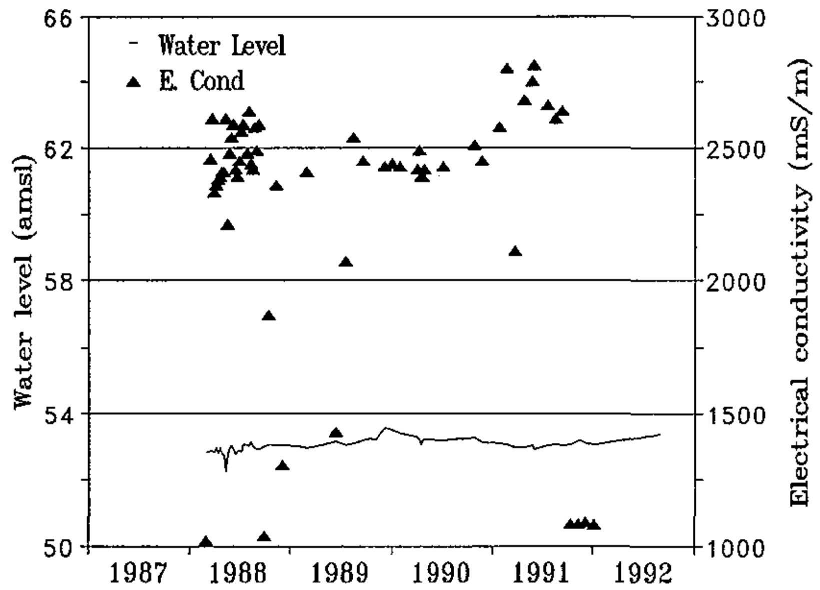


Figure 6.13 Water level and electrical conductivity at borehole CRG71

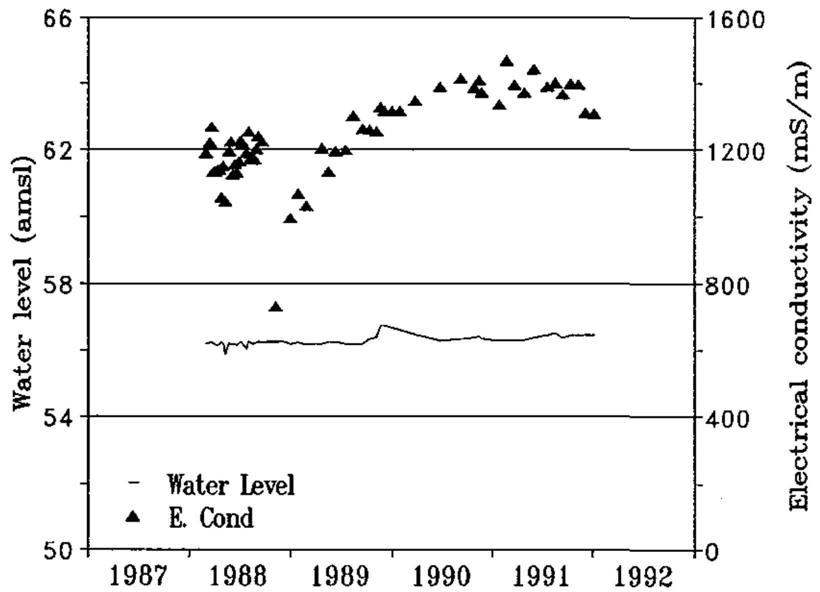


Figure 6.14 Water level and electrical conductivity at borehole CRG69.

At CRG71, the borehole located at the centre of an area with a very high surface salinity, there is a rise from approximately 2500 mS.m^{-1} to nearly 2800 mS.m^{-1} during the first half of 1991 (figure 6.13). However, the surface salinity at this site then drops to nearly 1100 mS.m^{-1}. This dramatic change in salinity is also difficult to explain without information on

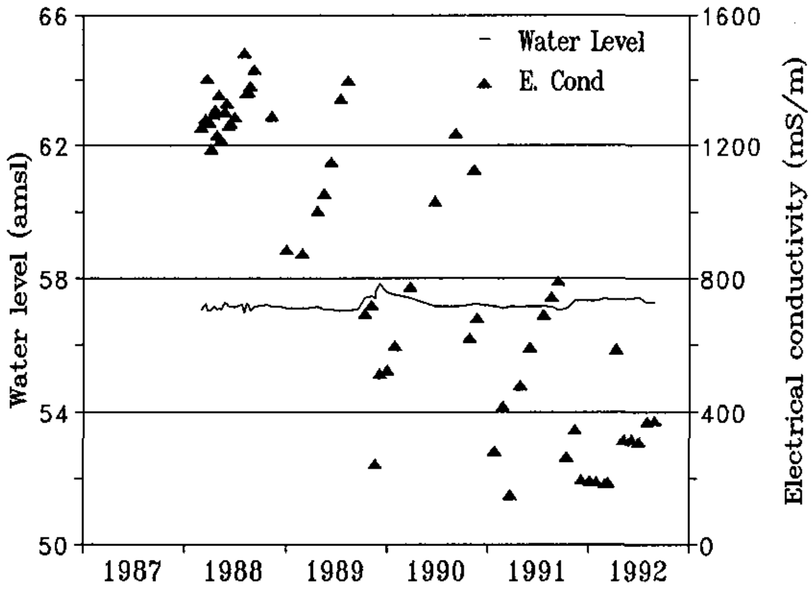


Figure 6.15 Water level and electrical conductivity at borehole CRG68.

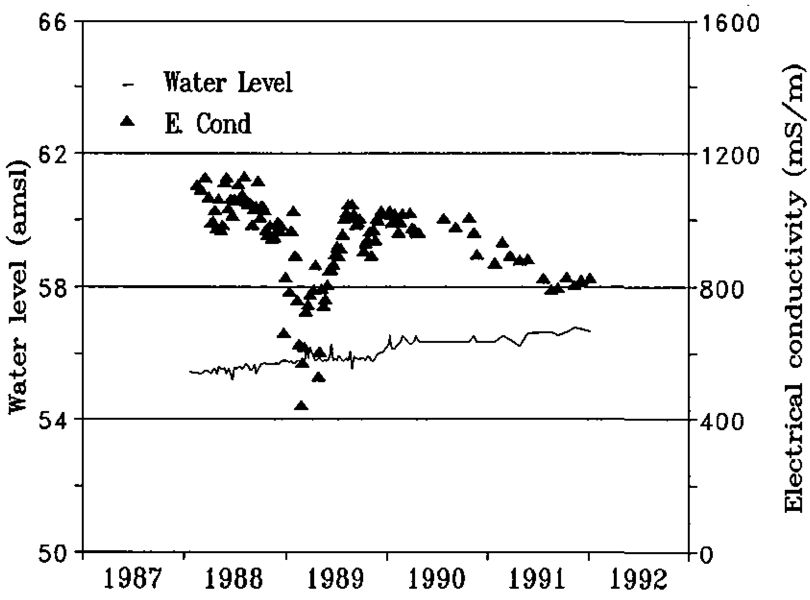


Figure 6.16 Water level and electrical conductivity at borehole CRG76.

the salinity depth profile. However, the probable cause is the simultaneous reduction in the rate at which very saline artesian ground water is moving up through the cretaceous mudstones from the underlying Table Mountain Sandstones as well as an increase in recharge from less saline irrigation drainage. The current prolonged period of drought will have lead to a drop in level of the regional ground water table and hence a reduction in piezometric head of the aquifer below the cretaceous mudstones of the Sundays River and Kirkwood formations. Therefore the rate at which the alluvial aquifer is recharged by the very high saline water moving up through weaknesses in the mudstone aquiclude is reduced. Concurrently, irrigation farming within this area of the Coerney valley is expanding, resulting in an increase in the recharge of the alluvial aquifer from irrigation drainage. This is clearly shown by the rise in water table at CRG71. It would therefore seem that the very saline ground water at this site no longer extends to the surface of the aquifer. A layer of less saline water comprising a mixture of very saline natural ground water and the less saline irrigation drainage water now forms an upper layer of the aquifer at this point.

The salinity of the aquifer at CRG69 (figure 6.15) steadily increases from the beginning of 1989 to a peak towards the middle of 1991. Subsequently, there is a small decrease although this new trend is as yet unclear. The peak salinity at CRG69 coincides with the peak levels recorded for CRG71, CRG81 and CRG82.

The salinity of the upper ground water collected at CRG68 (figure 6.15), which is located near the centre of a large field irrigated with a centre pivot overhead spray system, displays a great deal of variation. However there is a general trend of decreasing salinity. The variability of the sampled waters is attributed to the shallow water table which is only 1.2 metres below the surface and the extremely high rates of irrigation regularly applied. Unfortunately, the farmer of this land keeps very few records and is unable to provide information on the actual amount applied. However, the field is permanently wet and at times difficult to move across on either foot or vehicle. Currently the Department of Agriculture is designing a drainage system for this land. The variability is therefore related to the residence time of the surface water and the degree to which the most recently applied water mixes with the ambient ground water. The general decline in the salinity levels may relate to leaching of these soils with irrigation water.

The time series plots for CRG76 also display considerable variations through time. However, there is no simple explanation (figure 6.16). It is interesting to note that subsequent to a drop in salinity during the latter half of 1988 and first few months of 1989, the level of salinity rapidly rises during a period when the ground water level remains fairly stable. Then with the major rainfall and recharge event of November 1989 the water level rapidly rose and salinity decreased, albeit for only a few months. Subsequently the water table steadily rose while the water salinity dropped from 1000 to 800 $mS.m^{-1}$. It is interesting that the groundwater salinity changed so markedly, yet the lands surrounding CRG76 were not developed for irrigation until early 1992. It is therefore difficult to provide an explanation for the variations at this site.

Near the lower end of the valley the water level as recorded at CRG76 has remained relatively constant for the duration of the study period. Yet during this time the salinity of the surface water has increased from 1000 to nearly 1600 $mS.m^{-1}$. This increase in salinity coincides with increases in monitored at many sites within the valley including that at CRG81 which is well up valley and above the area currently influenced by irrigation drainage.

At site CRG93, near CRG82, an old disused borehole (depth of 350 metres) has also been sampled. This hole produced good quality artesian flow until the mid 1980's when its casing was corroded at the level where it passes through the saline water of the alluvium. Water from this hole is now contaminated by the saline water. Currently the surface water within this hole has a salinity of 150 $mS.m^{-1}$ and its level is within 1 metre of the ground surface. Before the corrosion of the borehole casing the water from this hole was used for drinking, without pumping, in a house with an elevation similar to that of the ground water at CRG83. Currently it is only possible to sample the surface water of the deep borehole near CRQ81.

6.7 INTERACTION BETWEEN THE SURFACE AND GROUNDWATER SYSTEMS

Before providing an overall description of the ground water system within the lower Coerney valley the study takes a closer look at the interaction between the alluvial aquifer and the Coerney River. Along the left bank of the Coerney River between flow monitoring sites CRG07 and CRQ08 there are a number of relatively strong seepages flowing into the river. One of these seepages, CRQ20 has been regularly monitored and found to have a salinity of approximately 2000 mSm^{-1}. As already discussed in this chapter it appears that at a site immediately up valley from this section of channel the alluvial aquifer is recharged by very saline artesian water leaching up through the underlying cretaceous mudstone. Some of this water then seeps into the stream channel raising the salinity of the streamflow at this point to over 1000 mSm^{-1}. Adjacent to this area, on the right bank of the river a number of shallow boreholes were drilled through the alluvium and into the first two metres of the cretaceous mudstone. The level and salinity of the water within these boreholes were also monitored regularly.

These water levels and that of the river clearly indicate a hydraulic gradient away from the right bank of the stream channel. Therefore over a relatively small area of the valley one can observe highly saline water rising up into the alluvium from the underlying mudstones. Some of this water then drains as bank seepages into the river on its left bank. Simultaneously, as indicated by the hydraulic gradient, water infiltrates from the right bank of the river into the ground water system. In other words the river at this point is both influent, on its left bank, and effluent on its right bank. It was possible to determine the rates of these seepage flows by solving the mass balance equation for this stretch of river between CRQ07 and CRQ08 using representative values of discharge monitored at these two water quality monitoring sites and salinity values of the influent and effluent seepages. By this method it was estimated that nearly 7 $l.s^{-1}$ of very saline water entered the river as seepages on the left bank and nearly 3 $l.s^{-1}$ of water drained into the alluvium between CRQ07 and CRQ08. These flow rates may not seem significant in terms of the total volume of water moving through the alluvial aquifer and streamflow system but they do indicate the complexity of the system in terms of both water movement and salinity mixing. A conventional water quality investigation of this aquifer provided little assistance in gaining an understanding of this hydrosalinity system. However, as discussed in Chapter 7 a more clear understanding of the interaction between the ground and surface water bodies within the lower Coerney valley is gained by examining the stable isotope composition of these different waters.

6.8 THE GROUND WATER SYSTEM OF THE LOWER COERNEY VALLEY

Consideration of the spatial variations in both the ground water level and its salinity as well as the salinity profiles down respective boreholes and the observation of major seepages flowing into the Coerney River provides a reasonably good conceptual understanding of the ground water system within the lower Coerney valley. It would appear that there are three major inputs to the system. Firstly, recharge within the lower Coerney valley itself with drainage from irrigation being a major component. Secondly, the ground water that flows through the alluvial aquifer from further up the Coerney valley. The third source is the very saline water rising up through the mudstones from a deeper confined aquifer. This aquifer is likely to be within the Table Mountain Sandstone, now overlay by the Kirkwood and Sundays River formations. In the past, farmers near the northern extreme of the study area used water of relatively low salinity from two deep boreholes (300-350m deep). These holes, which were drilled at least thirty years ago, are both artesian and flowed at the ground surface when initially drilled. Unfortunately, there are no logs for these holes and so there is little information available on the material through which they are drilled or of the deeper aquifer other than that it must be confined by the over lying mudstones. This conceptual understanding is supported by the isotope study described in Chapter 7 of this report.

The salinity of the Coerney River rises rapidly between water quality monitoring sites CRQ08 and CRQ12 (figure 7.13). Along the left bank of this section of the river there are numerous saline seepages which flow continually. The salinity of a one representative seepage (CRQ20), which has been monitored on a weekly basis, ranges from 1800 to 2000 mSm^{-1}. Near by, a shallow borehole penetrating the mudstone by less than two metres produced an artesian flow with a head of approximately 2.5 metres above ground level (57 m above mean sea level). Unfortunately, only one water sample was collected from this flow before the driller moved the casing and disturbed the base of the hole. The salinity of this water was approximately 1500 $mS.m^{-1}$, but it may have been diluted by the water used to lubricate the drill. It seems that an aquifer in the Table Mountain sandstone is confined by the over lying mudstone and siltstone deposits of which the Sundays Formation is marine laid and therefore highly saline. Water moving up through the mudstone becomes saline and provides a major source of salinity into the lower Coerney River.

7. STREAMFLOW WITHIN THE LOWER COERNEY VALLEY

The lower Coerney River is perennial from a point between river monitoring sites CRQ09 and CRQ10 (figure 7.1). Only once since January 1988 has the river flowed from above CRQ11. This was subsequent to the event of November 15, 1989. The river displays a very unresponsive flow regime and is generally unresponsive to all but exceptionally large rainfall events. It is therefore thought to be largely regulated by the local ground water system. The local farming community report that through time the extent of the perennial reach of the river has extended up stream. This may suggest a general rise in the water table as a result of the increased irrigation within the lower Coerney River valley. This is further supported by the steady increase in the base flow regime of this river during the study period.

Within the lower Coerney valley streamflow was continuously monitored at three sites: CRQ03, CRQ08 and CRQ11 (figure 7.1). Water samples were collected on a daily bases at CRQ03 and CRQ08 using pump samplers and manually at a further seven sites at weekly intervals. Two flow gauging surveys have also been carried out to determine the streamflow and salinity load at each water quality monitoring site where discharge was not continuously monitored. The results of this monitoring programme have provided information on variations of discharge and salinity through time as well as providing discharge and salinity profiles for base flow conditions along the entire length of the perenial river.

7.1 STREAM DISCHARGE OF THE LOWER COERNEY RIVER

Under base flow conditions streamflow within the Coerney River first emerges between sites CRQ09 and CRQ10. However, it is not until the river has passed site CRQ12 that the flow begins to rapidly increase from approximately 10 $l.s^{-1}$ to nearly 60 $l.s^{-1}$ at CRQ03 (figure 7.2). This increase in flow coincides with a section of channel where its elevation decreases as it descends from the Addo to the Colchester terrace and could be related to groundwater levels in the adjacent alluvium.

Figure 7.3 shows a significant increase in base flow discharge since monitoring began in mid 1987 which coincides with a significant increase in irrigation within the valley. A record of the quantity and spatial distribution of water application has been compiled from records of the Lower Sundays River Irrigation Board. It should be noted that most of the irrigated lands developed since the beginning of this study have been upstream of flow monitoring site CRQ08. Yet the increased flow at CRQ03 since 1987 is an order of magnitude greater than the normal base flow measured at CRQ08. Perhaps this highlights the importance of understanding the groundwater system of the lower Coerney valley before attempting to construct a water or salt budget for the area.

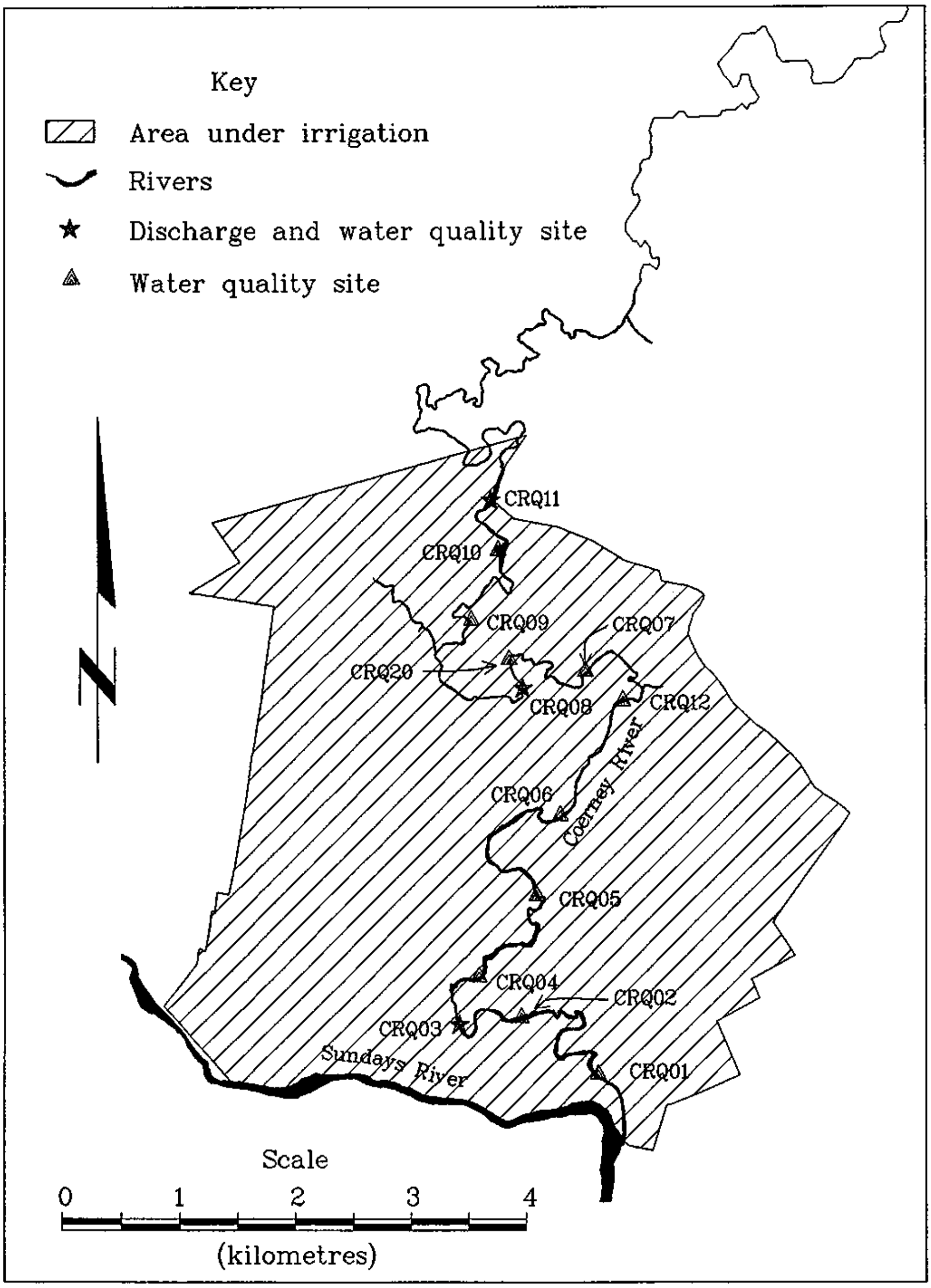


Figure 7.1 Streamflow and water quality monitoring sites on the Coerney River.

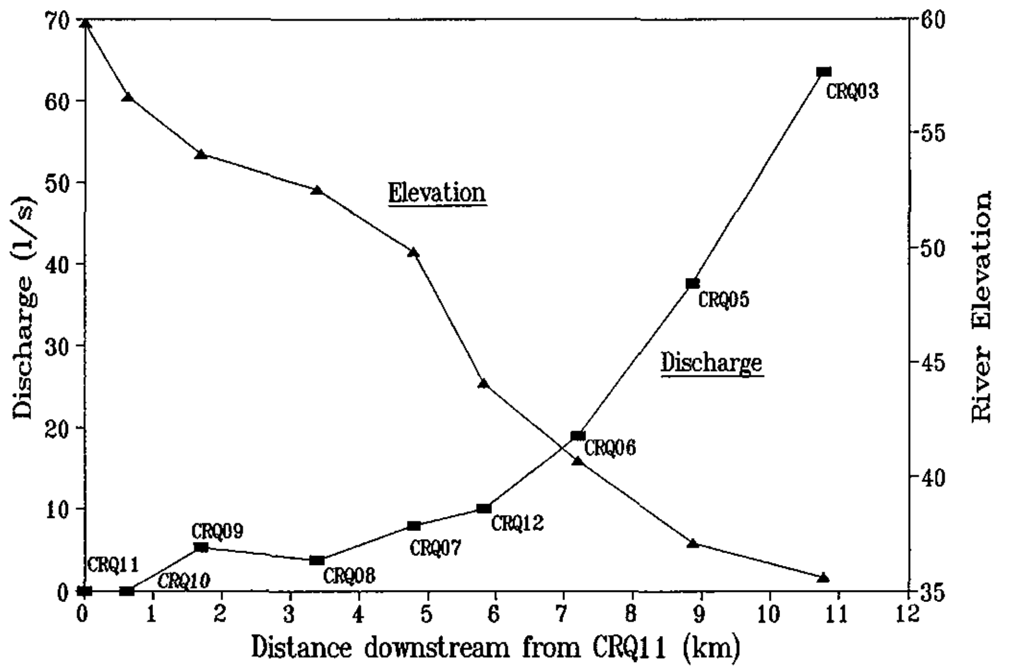


Figure 7.2 Profiles of channel elevation and base flow discharge down the Coerney River.

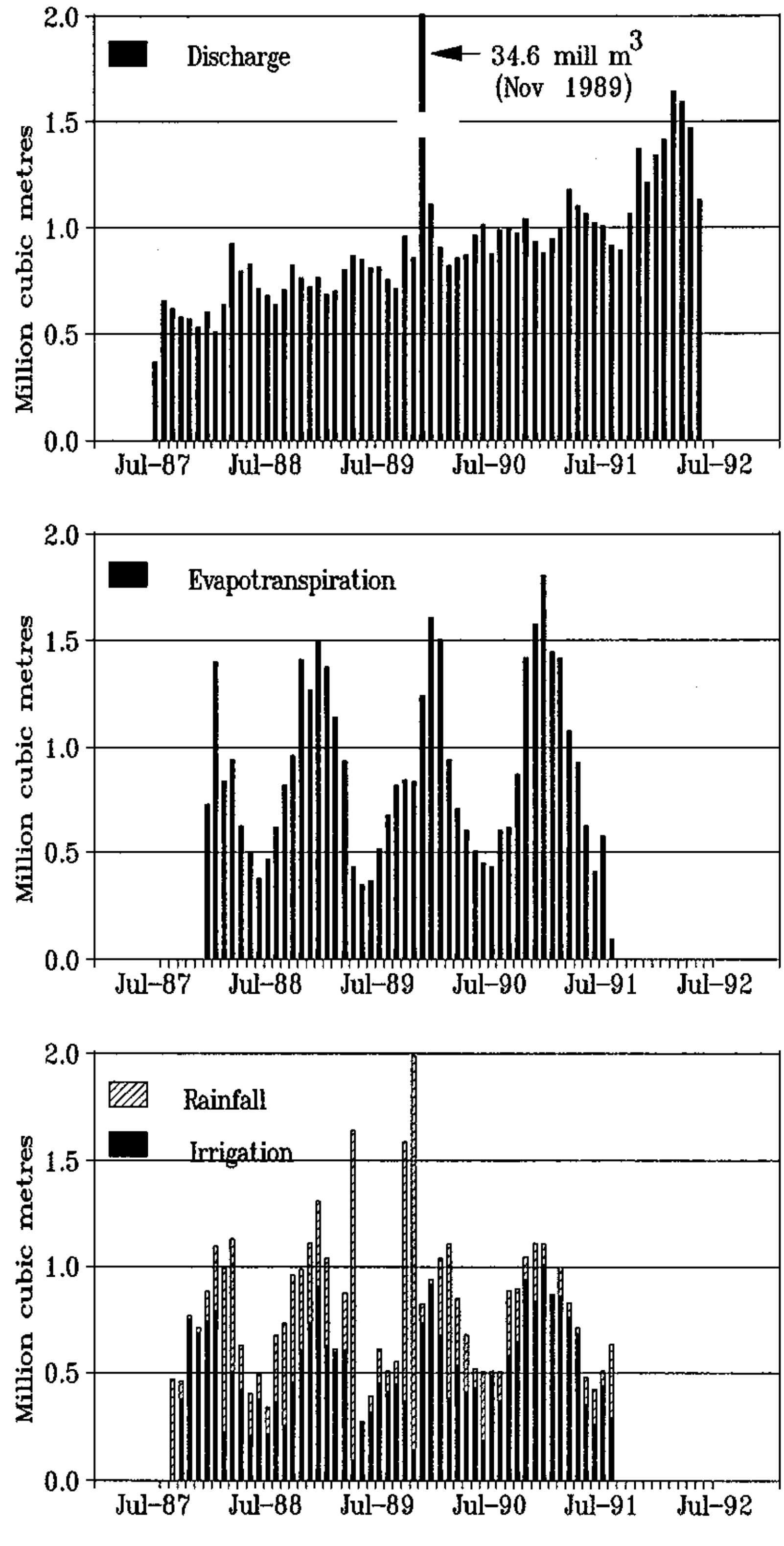


Figure 7.3 Variations in the irrigation water supplied to the Lower Coerney valley and stream discharge measured at CRQ03.

7.2 SALINITY OF THE LOWER COERNEY RIVER

Within the Lower Coerney River there are two distinct sources of salinity. Firstly, the geology of the study area is largely comprised of marine laid sediments which tend to leach salts via the ground water system into surface streams. Secondly, with the high rates of evaporation and low rainfall characteristics of this area, salts tend to accumulate within the soil. These salts may be flushed into the river system during major rainfall events. This process may be inhanced by the addition of irrigation water, especially if the applied water has a high salt content. This section of the study sets out to examine the salinity of the lower Coerney River and determine the contribution of salts from these different salinization processes. For the purpose of this study salinity is expressed as both electrical conductivity (EC) in $mS.m^{-1}$ and as total desolved solids (TDS) $mg.l^{-1}$. The relationship between TDS and EC values as recorded at sites CRQ03 and CRQ08 was determined using regression analysis and is plotted in figure 7.4. This enables the ready conversion of EC to TDS to provide common units for comparison.

Typical variations in salinity down the Lower Coerney River are shown in figure 7.5. From it emergence upstream of CRQ09 the TDS is shown to increase slowly to approximately 5000 $mg.l^{-1}$ at CRQ08. From this site the salinity of the river increases very rapidly to over 7500 $mg.l^{-1}$ at CRQ12. This increase is attributed to the inflow of saline seepages of over 11000 $mg.l^{-1}$ which are clearly observed on the left bank of this reach of the river. Water samples from one of these seepages, CRQ20, was monitored on a weekly basis. In Chapter 6 of this report the rate of seepage inflows between sites CRQ08 and CRQ07 are estimated to be 7 $l.s^{-1}$. This saline water is thought to rise up through the marine laid Sundays River formation which acts as an aquiclude from the regional groundwater system of the underlying Table Mountain Sandstone. Downstream of site CRQ12 the salinity of the Coerney River decreases to approximately 4500 $mg.l^{-1}$ at CRQ03. This decrease is attributed to the dilution of the saline streamflow by less saline irrigation return flow waters. However, due to the increasing rate of flow the salt load below site CRQ8 increases significantly (figure 7.6). Currently the base flow salt load at CRQ08 is estimated to be approximately 2 tonnes per day while that at CRQ03 the load is nearly 24 tonnes per day.

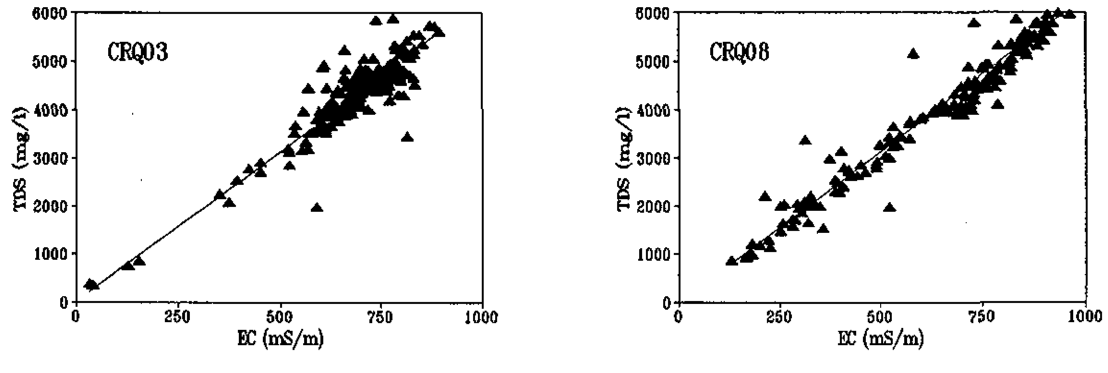


Figure 7.4 The relationship between electrical conductivity and total dissolved solids at sites CRQ03 and CRQ08 on the lower Coerney River.

During the period of the study there has been a decrease in the salinity of the base flow of the Coerney River at CRQ03 (figure 7.7). This is attributed to the increasing irrigation return flow component of this flow regime which, being of lower salinity than the saline water emerging from the mudstone beneath the alluvium, has a diluting effect. However, during this time the increased discharge, of albeit less saline water, has resulted in a doubling of the salt load of the Coerney River since July 1987. This is more clearly displayed by

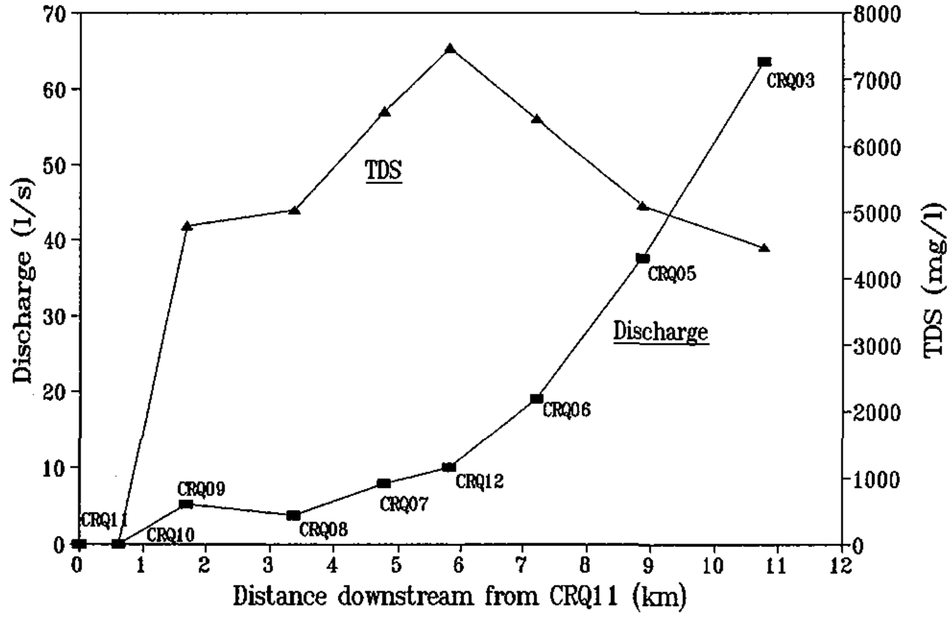


Figure 7.5 Variations of TDS concentration and discharge down the Lower Coerney River.

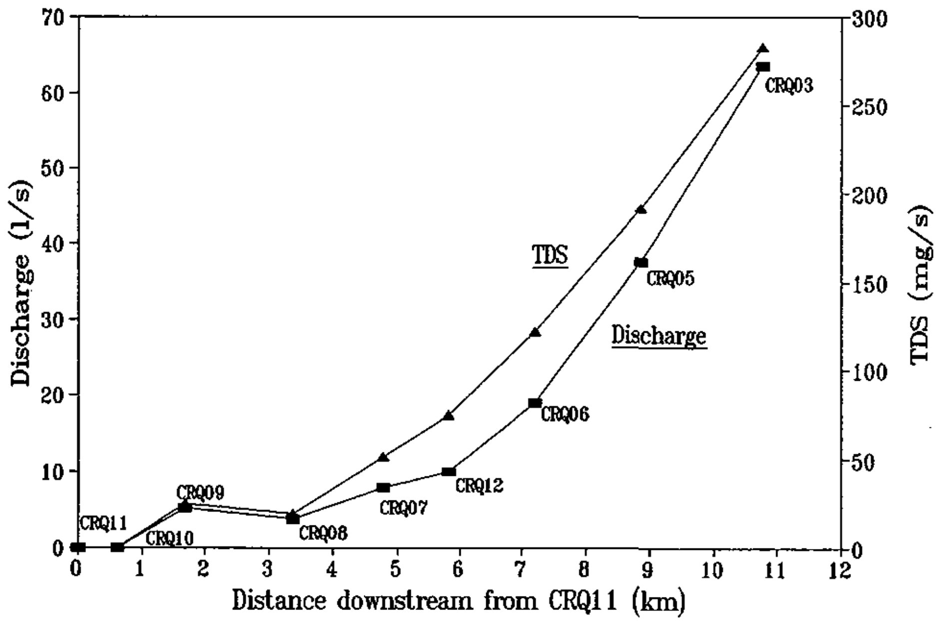


Figure 7.6 Variations of TDS load and discharge down the Lower Coerney River.

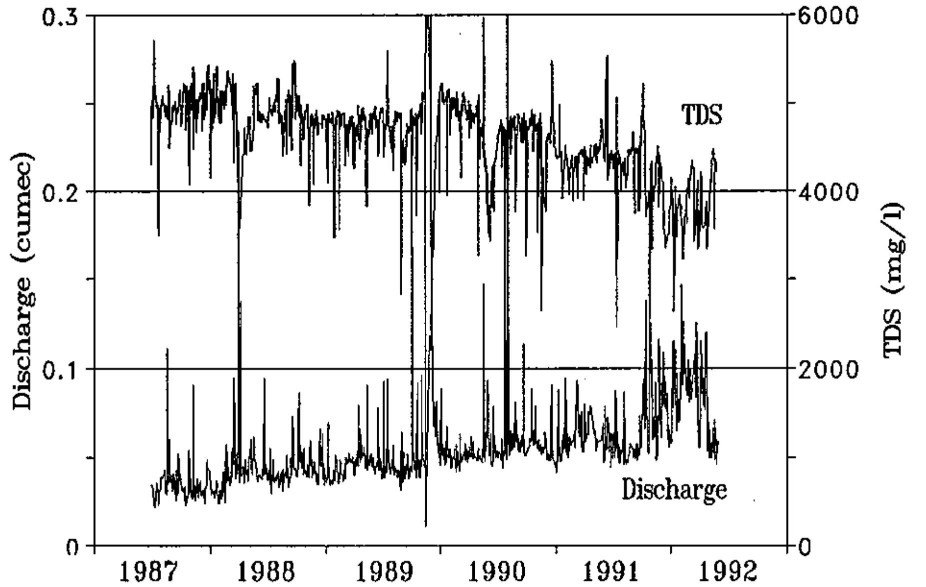


Figure 7.7 Variations of TDS and discharge of Coerney River at CRQ03.

figure 7.8 which presents monthly salt loads estimated for the Coerney River at CRQ03. It should be noted that the estimated salt loads are for the Coerney River at CRQ03 and underestimate the total salt load entering the Sundays River from the Coerney Valley due to the flow of saline water through the alluvial system. To more accurately determine a salt budget for the lower Coerney valley, rates of groundwater flow will being determined by modelling the groundwater system.

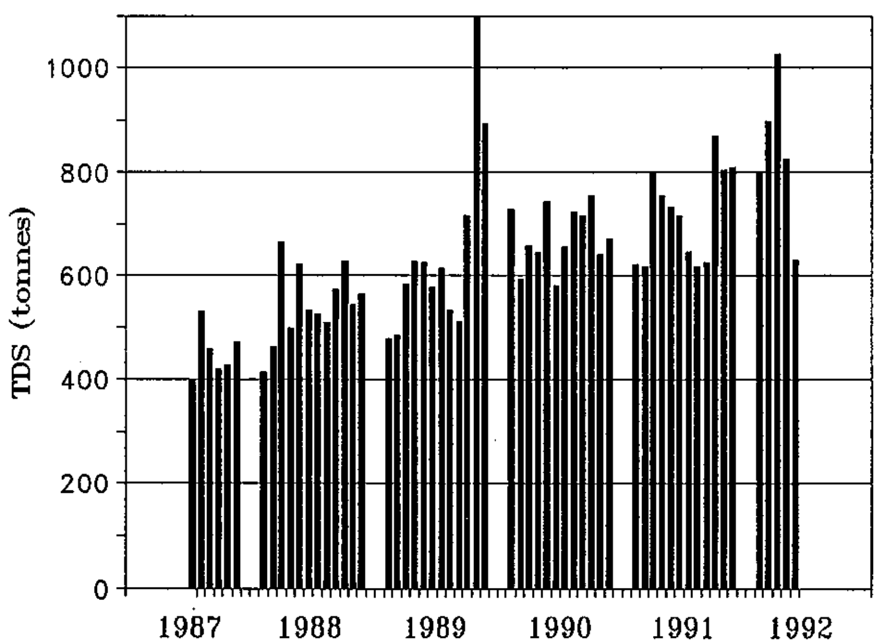


Figure 7.8 Monthly salt loads of the Coerney River at site CRQ03 since July 1987

7.3 SOURCES OF SALINITY WITHIN THE LOWER COERNEY VALLEY

It is clear that the base flow discharge and solute load of the Lower Coerney river have increased significantly during the study period (figures 7.7 and 7.8). These changes can only be attributed to the increasing area of irrigation. The proportion of the Coerney's streamflow, comprising irrigation return flow, is determined by analysing the relative concentrations of the stable isotopes ^2H and O^{18} at the different sampling sites.

Due to preferential precipitation, the atmospheric moisture of an air mass becomes increasingly depleted of the ^2H isotope. This is reflected by the relative depletion of ^2H in precipitation which falls inland away from its maritime source. The isotopic relationship of ^2H to O^{18} for samples of rainfall collected within South Africa is defined by the SA meteoric line as displayed on figure 6.5. Subsequent to precipitation, evaporation leads to further enrichment of the heavier O^{18} isotope. As stable isotopes are conservative, an analysis of their relative concentration not only provide a good indication of a waters evaporative history, but can also provide information on its mixing with other waters of known isotopic concentrations.

The irrigation water used in the lower Coerney Valley is brought some 600 km from the Hendrik Verwoerd Dam, on the Orange River, by a system of canals, controlled river channels and balancing dams. This prolonged period of exposure to evaporation leads to its relative enrichment with the heavier O^{18} isotope as is clearly shown by the plotting position of the canal sample SNC10 on figure 7.9. The O^{18} enrichment process continues as the water applied to the citrus orchards within the valley is further evaporated. This is shown by a further shift to the right of the plotting position for water collected from river site CRQ08. In comparison, the deep ground water from the sandstones of the Cape Super Group, sampled from the artesian flow of borehole CRG93, plots on the lower end of the SA meteoric line. This isotopic relationship suggests that the deep ground water has not been exposed to excessive evaporation and has not been depleted of ^2H. If one assumes that streamflow water sampled at CRQ08 is primarily from irrigation return flow and that ground water sampled from CRG81, which is well outside the area of irrigation, represents the natural mix of the deep ground water and locally recharged precipitation, then it is possible to apply a simple mixing model to determine the changing composition of water down the Coerney River.

$$IR = \frac{(C_m - C_{gw})}{(C_{rf} - C_{gw})} * 100 \qquad \text{eq.1}$$

where IR = percentage irrigation return flow
C_m = O^{18} conc. of irrigation and ground water mix
C_{gw} = O^{18} conc. of natural ground water
C_{rf} = O^{18} conc. of irrigation return flow

The mixing model shows that under current baseflow conditions, 69 % of the flow at CRQ03 comprises irrigation return flow water (Table 7.1). Table 7.1 also shows that towards the outlet of the catchment the shallow alluvial ground water aquifer is comprised of an increasing proportion of irrigation water. It is reasonable to assume that the leaching fraction of the areas currently being developed for irrigation will be similar to those areas already in

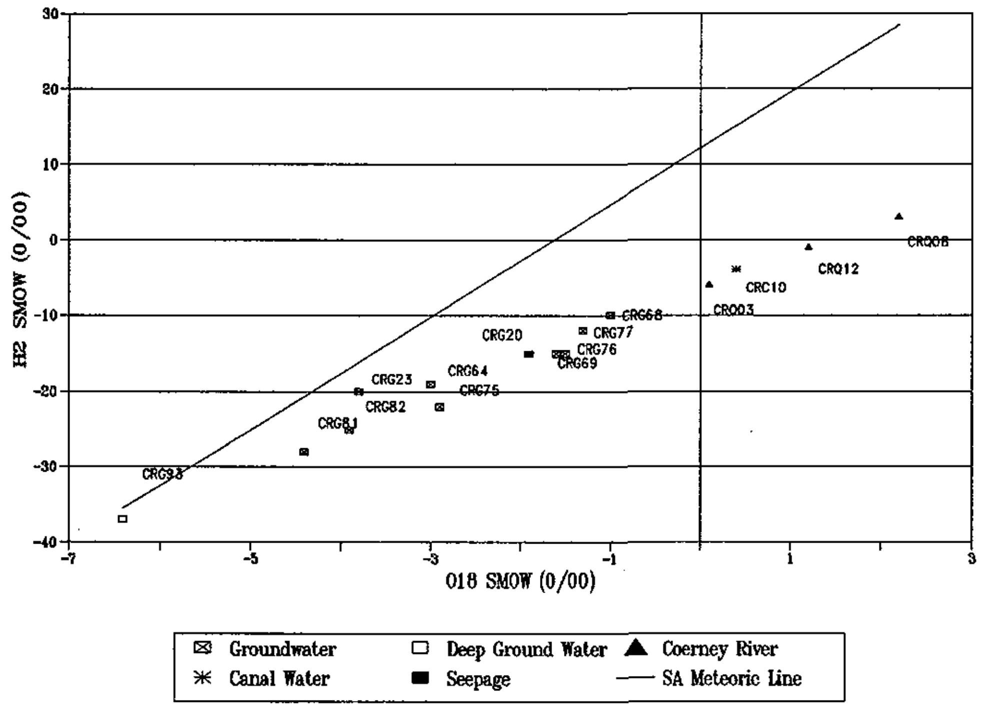


Figure 7.9 Isotopic relationship of different waters within the lower Coerney Valley

Table 7.1 Stable isotope concentrations and the composition of water samples collected at different sampling sites.

Sampling Site	^{2}H (SMOW)	O^{18} (SMOW)	% water from Irrigation
Streamflow			
CRQ08	3	2.2	100
CRQ12	-1	1.2	85
CRQ03	-6	0.1	69
Bank Seepage	-15	-1.9	38
CRQ20			
Ground Water			
CRG23	-18	-3.4	9
CRG64	-26	-3.7	21
CRG68	-10	-1.0	52
CRG69	-15	-1.6	42
CRG75	-22	-2.9	23
CRG76	-15	-1.5	44
CRG77	-12	-1.3	47
CRG81	-28	-4.4	0
CRG82	-25	-3.9	8

production. Therefore, the flow of the Coerney River is expected to increase due to additional irrigation return flows. This will further dilute the highly saline natural ground water of the alluvium and lead to a reduction in the salinity of the lower Coerney River. However, the solute load of the river will continue to rise with the increased volume of streamflow.

7.4 STREAMFLOW SALINITY AND DISCHARGE REGIMES

Figure 7.8 is dominated by the salt load of November 1989. During this month there was a major storm event which unfortunately exceeded the capacity of the flow measuring structures on the Coerney River. It was also unfortunate that the pump sampler at CRQ03 had to be removed to prevent it from being washed away. However, by simply extrapolating the rating curves for the respective flow monitoring sites, estimates of the discharges during this event have been determined. This rating curve extrapolation does not consider an increase in the channel cross-sectional area with increasing stage and therefore provides very conservative estimates of the stormflow discharges. Water samples were collected manually during this event at CRQ03 during the period when the pump sampler was removed to safety.

The salinity and estimated flow rates for this event at CRQ08 are presented in figure 7.11. It will be noted that during the rapid rise in discharge the salinity decreased rapidly from

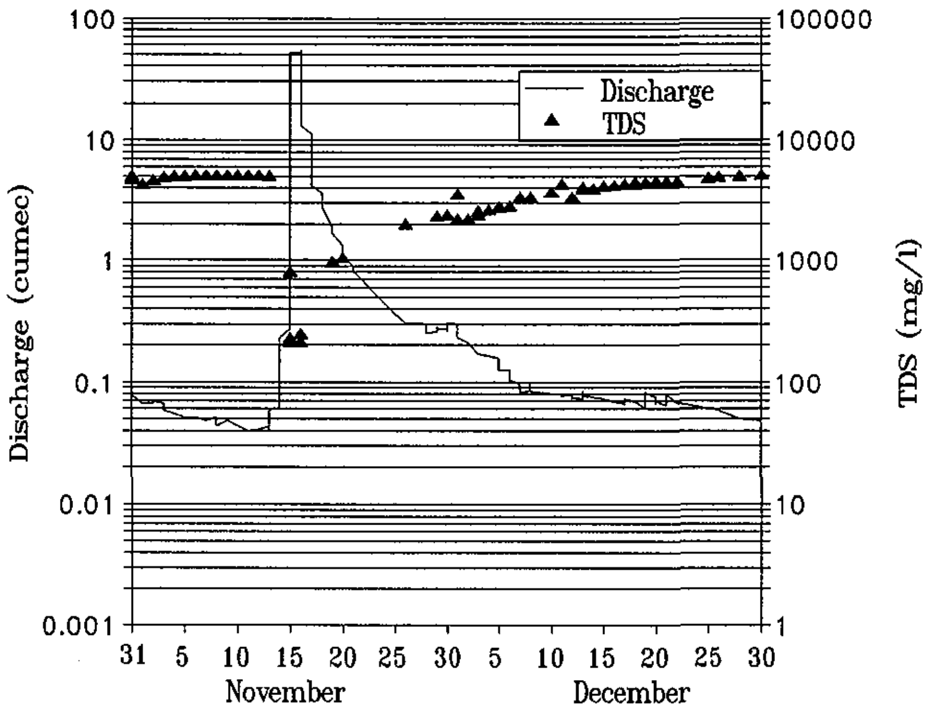


Figure 7.10 Discharge and salinity of the Coerney River at CRQ03 during the flood event of November 1989.

approximately 5000 mg.l^{-1} to less than 200 mg.l^{-1}. Subsequently the salinity increased to a level greater than that before the storm event and then stabilized again at approximately 5000 mg.l^{-1}. This suggests a process of initial dilution by the increasing stormflow with water flowing into the alluvium from the increased river stage. As the stormflow receded the salinity of the river increased as water flowed back from the alluvium into the river system.

The mobilisation of salts stored in the alluvium in close proximity to the channel resulted in a rapid rise in streamflow salinity during the recession of this storm event.
The total salt load estimated for the Coerney at CRQ03 during November 1989, which was dominated by this flood event, is approximately 2500 tonnes. Unfortunately, as only a crude estimate of the stormflow hydrographs for the flow monitoring sites on the Coerney can be constructed, it is not be possible to determine the salt load originating from with the irrigated area of the catchment.

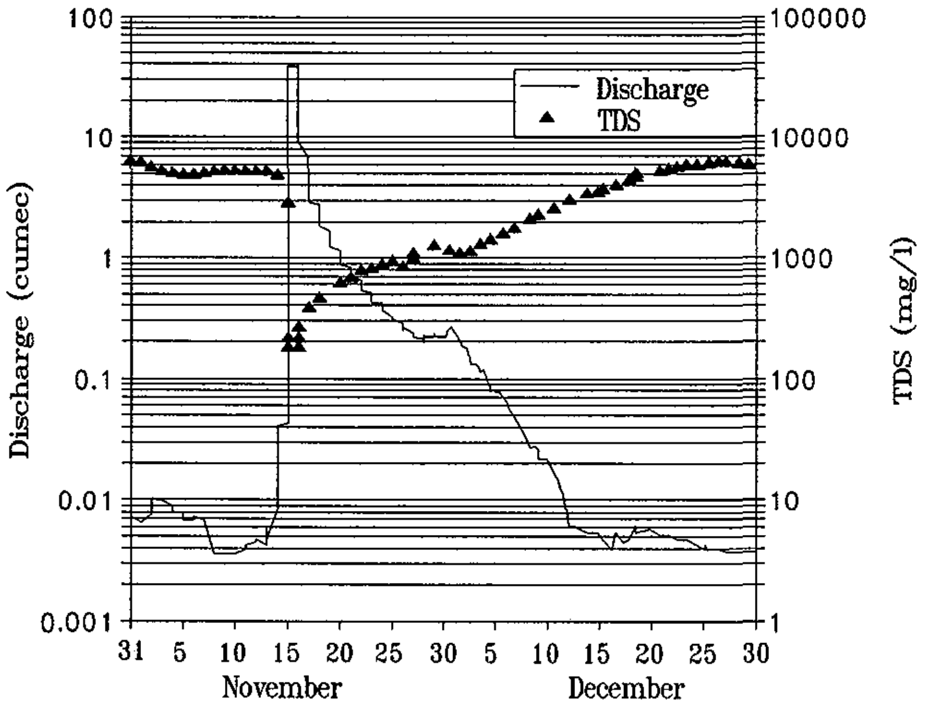


Figure 7.11 Discharge and salinity of the Coerney River at CRQ08 during the flood event of November 1989.

8. CONCLUSIONS AND RECOMMENDATIONS FOR FUTURE RESEARCH

The primary aim of this research was to gain a better understanding of the hydrosalinity processes operating within the lower Sundays River irrigation area. Within the project proposal the objectives of this study were split into four related components and a proposed methodology was described. However, as discussed in section 1.1 of this volume, these objectives and the proposed methodology were identified to be somewhat over ambitious and based on several misconceptions. Therefore to provide a more clear framework the following four more realistic objectives were identified:

i. To acquire data on the 3-D processes of moisture and solute movement at various spatial and temporal scales.

ii. To investigate and conceptually describe the hydrochemical processes within the root and delivery zone.

iii. To test components of selected root zone and ground water models at various spatial and temporal scales.

iv. To determine the impact of irrigation on streamflow and ground water within an area of irrigation.

The first two of these objectives are very similar to the first two objectives of the project proposal. However, the third and fourth objectives of the proposal, to test and combine components of root zone and ground water hydrosalinity models, and to develop a simplified model for irrigation management were clearly over ambitious given the resources and time available. Also, the original proposal does not specify determining the impact of the irrigation on the surface and ground water systems of an area under irrigation as a specific objective. This has been included as the fourth revised objective for this study.

To meet these objectives instrumentation was installed and an intensive data collection programme carried out. The following discussion will highlight specific conclusions reached in satisfying each of the four main revised objectives. Also, several specific recommendations for future research are proposed.

8.1 CONCLUSIONS

Initially the study examined variations of soil water content, and soil water and solute flux within the root zone. Data were collected at a number of micro-plots, but only that for one plot, DDM03, located in orchard M of Daisy Dell farm was used for this investigation. The most important conclusion drawn from this aspect of the study was the importance of macro-pore flow. Soil moisture flux was initially determined using a Darcian approach which is driven by the difference in matrix potential between two points in the soil profile. This method, which can only account for micro-pore flow, was found to grossly underestimate the total soil water drainage. A water balance approach was found to provide more realistic results, but was discarded due to its gross over estimation of negative fluxes when the evaporative term exceeded available soil moisture. The study clearly demonstrates how the water balance approach can not be implemented using a fixed crop factor for determining the evaporative term. To more successfully implement this approach one would require the knowledge needed to more accurately define the transpiration response of citrus trees to variations in the factors controlling the evaporative processes. Therefore, due to problems related to both the Darcian and the water balance methods of determining irrigation drainage, a third approach based on a chloride mass balance was developed. This approach is believed to provide the most realistic estimate of soil moisture flux. The soil moisture fluxes determined by this approach were very similar to those determined by the water balance approach, but without the large negative fluxes between irrigation events.

A comparison of the soil moisture fluxes as determined using the Darcian approach, which can only account for micro-pore flow, and that determined using the mass balance approach, clearly indicated the importance of macro-pore flow as a component of irrigation drainage. The chloride mass balance approach was subsequently used to determine the solute flux within the root zone of the study orchard.

The third objective of the study was to evaluate a number of root zone models. Initially it was suggested, by the Steering Committee for this project, that this aspect of the study should rely on the findings of a another Water Research Commission funded project which set out to evaluate a number of hydrosalinity models for the specific purpose of estimating soil water and solute movement within the root zone (Moolman, 1993). However, as the progress of that study was delayed, this project undertook a more independent approach. Three management and one research type model were selected to cover the full range of model complexities. The output from all four models, when compared with the soil moisture fluxes determined by the mass balance approach, were very disappointing. These results are of great significance as they clearly demonstrate the need for hydrosalinity models that are applicable in soils where macro-pore flow is the dominant form of soil moisture drainage. This is an important drainage process, especially where farmers are encouraged to over irrigate by the relatively low cost of irrigation water. Within South Africa, which is faced with ever increasing water resource limitations and where over 50 percent of this resource is used for irrigation agriculture, there is an urgent need for research into, and improved management of, irrigation farming. The development of applicable hydrosalinity research and irrigation management models is an area of research that should be given a far greater priority than is currently the situation.

The fourth objective of the study was to determine the impact of irrigation on both the ground water and streamflow within the study area. A number of boreholes were drilled by the Department of Water Affairs and Forestry. The level and salinity of the ground water aquifer within the alluvial deposit of the lower Coerney River valley were monitored at these sites. The vertical salinity profile of the aquifer was also recorded at each borehole. These data clearly indicated a marked rise in the level of the water table in response to irrigation farming. The area of influence extended some distance up valley from irrigation development. It was also learnt that the salinity of the irrigation drainage is significantly less than that of the natural ground water within the valley. An isolated pocket of highly saline water was very clearly indicated by constructing an isoline map of electrical conductivity values recorded within the lower Coerney River valley. The salinity of this water is very similar to that of the aquifer further up valley from the area influenced by irrigation. An examination of the vertical salinity profile at this point indicated that the highly saline water extended from the upper to the lower surface of the aquifer. In close proximity to this area of high salinity, a shallow borehole drilled through the alluvium and into the underlying cretaceous mudstones struck artesian water. This suggests that artesian water is rising up through the marine laid mudstones from the even deeper Table Mountain Sandstones. After moving up through the marine laid mudstones, the relatively less saline water of the Table Mountain Sandstone has an electrical conductivity in excess of 2000 $mS.m^{-1}$. The less saline irrigation drainage water initially floats above the more dense, very saline, natural ground water. The salinity of this aquifer is therefore highly variable with both lateral and vertical gradients. This understanding clearly indicated the dangers of monitoring the surface water of an aquifer in the hope of determining its water quality characteristics.

The study also examined the discharge and salinity regimes of the lower Coerney River itself. Discharge records clearly indicate a steady rise in the base flow of the river as irrigation farming expands within the valley. Simultaneously the solute concentration of the river has decreased as the natural ground water component of the streamflow is diluted with an increasing proportion of less saline irrigation return flow. Stable isotopes were successfully used to determine the natural ground water and irrigation return flow components of both streamflow and local ground water at various locations within the study area. This technique was found to provide more useful information than conventional water quality analyses for gaining an understanding of the hydrological and hydrochemical systems of the study area.

8.2 THE EXTENT TO WHICH PROPOSED OBJECTIVES WERE MET

As previously discussed, given the time and resources available, the specific objectives of this project were found to be overambitious. However, many of the objectives were met, and for those as yet unobtained the knowledge gained from this study provides a better conceptual understanding of the system such that these objectives may be satisfied in the future.

The first objective, to investigate and describe the hydrochemical processes within the delivery zone was not fully achieved. It was initially proposed that the study should monitor the inputs and outputs for representative cells (1 - 10 km^2) within the alluvial valley of the study area. However, given the heterogeneity of the hydraulic characteristics of an alluvial deposit and the spatial and temporal variability of the land use management within such cells, this methodology is most inappropriate. As such cells can not be defined as water tight units it is not even possible to construct crude mass balances which can be related to variations in the land use practice applied to their surface. Clearly, at the time when the proposal for this project was developed the conceptual understanding of alluvial and ground water systems and of irrigation return flow processes were overly simplified. These problems were further compounded by recharge of the local alluvial aquifer by highly saline artesian water rising up through the underlying mudstones. However, through the improved understanding of the hydrosalinity system gained by this study, and by the use of stable isotope analyses, which were successfully applied in this study, it should now be possible to formulate a more appropriate methodology to satisfy this objective.

The second objective was satisfied by establishing a number of micro-plots and monitoring both soil water and solute movement within the root zone. These data were used to examine a number of methods for estimating soil water drainage and to assess four root zone hydrosalinity models.

The third objective of the study was to assess hydrosalinity root zone and ground water models to identify model components that could be combined to form an irrigation return flow model for the study area. As the research proposal did not specify the specific root zone and ground water models to be assessed the Steering Committee recommended that this project should utilize the results of another Water Research Commission funded project that set out to evaluate a number of hydrosalinity root zone models (Moolman 1993). However, as progress from that project was delayed a decision was made to independently select and assess three management and one research level hydrosalinity model. These models were assessed using data from the micro-plots, established for this study, to determine their ability to predict soil water drainage. In conclusion, it was found that none of the models adequately predicted soil water drainage and that they could not be used as components for an irrigation return flow model.

The project also attempted to implement a 3-D finite difference ground water model to provide information on ground water flows within the alluvial deposit of the lower Coerney River valley. Although much of input data for the model was compiled, it could not be fully implemented in the time available owing to the unexpected complexity of the ground water system which is characterised by significant recharge from irrigation drainage. The distribution of this water is highly variable in both time and space. A further complication is the water rising up from through the underlying mudstones. The successful implementation of this model, at an appropriate spatial resolution, would provide further insight into this ground water system.

The fourth objective of the study was to develop a simplified model predicting irrigation returnflow at a catchment scale. As this was not possible with the time period of the project and also due to the disappointing performance of the models that were assessed as potential soil water drainage components for this model it was decided to implement an existing hydrosalinity systems model for the study area. The DISA model (Görgens, 1990), which was developed as an irrigation management systems model for the Breede River, was selected and the input data compiled. Unfortunately, owing to a programming error which the developers could not correct in time for the completion of this project, these data could not be entered into the models data base. However, the compilation of these data and through defining a model configuration for the lower Coerney River valley some aspects of this model have been assessed.

During the course of this study a modified list of objectives were defined to provide the project with a more clear research framework. The fourth of these objectives was to assess the impact of irrigation on the ground and surface ground water systems of an area under irrigation. In meeting this objective the study has gained a much greater understanding of the hydrosalinity system of the lower Sundays River valley and in particular that of the Coerney area. With the improved conceptual understanding and the establishment of a data base it should be possible to use this area to gain further understanding of the physical factors controlling the relevant hydrosalinity processes, and to further develop hydrosalinity and irrigation management models.

8.3 RECOMMENDATIONS FOR FUTURE RESEARCH

The current study was carried out at a number of scales, from micro-plot studies to an examination of the impact of irrigation on ground water and streamflow at the catchment scale. Analyses of the data collected for this study and conclusions drawn from an assessment of several hydrosalinity models has highlighted a number a areas for future research. In particular, it is suggested that there is a need for further understanding in the following facets of hydrosalinity research:

i. To establish a more practical methodology for determining the magnitude and spatial variability of macro-pore flow.

ii To carry out detailed studies of the hydrosalinity processes within the delivery zone at the plot scale rather than at the catchment scale.

iii. To gain a better understanding of the spatial variations of hydrosalinity processes at the field scale.

The aims of these studies should be to contribute to the further development of root zone and irrigation systems models that may be applied as management tools for the more efficient use of irrigation water and to lessen the impact of irrigation drainage on the limited water resources of Southern Africa. In particular it is considered important :

i. to develop an hydrosalinity research model that considers macro-pore flow and accounts for spatial variability at the field scale.

ii. to develop an irrigation systems model that permits the movement of ground water between adjacent ground water cells and both vertical salinity gradients within the local ground water system.

This study highlighted the importance of macro-pore flow as a dominant process in irrigation drainage. However, there is currently no sound methodology by which this component of soil water movement can be readily estimated and accounted for in hydrosalinity models.

Another very problematic area of salinity research lies with the lack of understanding of the processes operating within the delivery zone. This zone which connects the root zone to the ground water and surface water systems is very difficult to monitor because of its depth and also because it is often comprised of cobble and boulders which are difficult to penetrate with drilling equipment appropriate to research. Yet, being an important link for the transfer of irrigation drainage to the ground water and streamflow systems, it is important to understand how solutes may either be accumulated and leached from this zone if the impacts of an irrigation development are to be properly assessed.

For the successful extrapolation of hydrosalinity information from plot studies to the field or catchment scale it is be important to gain an understanding of the spatial variability of these processes. If one can statistically define these variations it may be possible to develop more stochastic type models and reduce the data requirements of spatially distributed, deterministic type hydrosalinity models. The potential advantages of this type of hydrosalinity modelling suggest that to further examine the spatial variability of hydrosalinity phenomena would be most worthwhile.

Currently, due to the increasing pressure on South Africa's water resources there is a great need for improved irrigation efficiency. If one considers the large volume of water currently used for irrigation, it is clear that even very small improvements in irrigation efficiency may lead to a relatively large saving in water. There is also an increasing need to reduce the pollution of especially surface water resources by irrigation drainage. For example, the salinity of water transferred via the Orange/Fish River scheme increases in salinity from approximately 35 $mS.m^{-1}$ at Katkop (Q1M01) to over 75 $mS.m^{-1}$ at Elands Drift (Q5R0101), a distance of less than 50 km. This increase in salinity is largely attributed to saline irrigation return flows generated along this reach of the Great Fish River. The cost of this irrigation practice in terms of loss of production in the lower Sundays River irrigation area has not been estimated, but it is most probably considerable. Appropriate hydrosalinity models would provide an ideal methodology for determining more efficient irrigation strategies with the potential of saving water and reducing pollution. However, before selecting a model for this work, it would be necessary to ensure that it was appropriate to the conditions of the study area. In particular, the importance of macro-pore flow must be determined and if necessary an appropriate model developed. Currently it would appear that such a model is not available and that research into this aspect of hydrosalinity modelling is essential.

To evaluate the impact of an irrigation development on either the ground or surface water resources of an area, an appropriate irrigation management systems model is required. Such a model has been developed for the Breede River (Görgens, 1990). However, as the structure of this model was developed specifically for the Breede River valley it does not facilitate the movement of ground water between adjacent cells. It also does not facilitate variations of salinity in the vertical profile of the aquifer and recharge of very saline water from an underlying artesian aquifer. It would seem that to meet these more flexible modelling requirements a finite difference type model would be required. Such a model would have wider application than the currently available model developed for the Breede River irrigation area.

Further to this list of specific hydrosalinity research requirements, it is recommended that the monitoring and analyses of data collected for the lower Coerney River valley, initiated in this study, should be continued. There is now an historical record of the impact of irrigation on both the ground water and streamflow systems within the valley. There is also a good conceptual understanding of the hydrological system and methods for identifying components of this system have been established. Currently, irrigation farming within the lower Coerney River valley is under rapid expansion following the construction of a new higher level canal. It would therefore make a great deal of sense for at least some of the recommended areas of research, as listed above, to be carried out in this study area.

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