The importance of aquitard windows in the development of alluvial groundwater systems : Lower Murrumbidgee, Australia

Author:
Timms, Wendy Amanda

Publication Date:
2001

DOI:
https://doi.org/10.26190/unsworks/20687

The importance of aquitard windows in development of alluvial groundwater systems: Lower Murrumbidgee, Australia.

W.A. Timms

A thesis submitted in fulfillment
Of the requirements for the degree of
Doctor of Philosophy

University of New South Wales
School of Civil and Environmental Engineering

2001

Disclaimer

I hereby declare that this submission is my own work and to the best of my knowledge it contains no material previously published or written by another person, nor material which to a substantial extent has been accepted for the award of any other degree or diploma at UNSW or any other educational institution, except where due acknowledgement is made in the thesis. Any contribution made to the research by others, with whom I have worked at UNSW or elsewhere, is explicitly acknowledged in the thesis.

I also declare that the intellectual content of this thesis is the product of my own work, except to the extent that assistance from others in the project's design and conception or in style, presentation and linguistic expression is acknowledged.

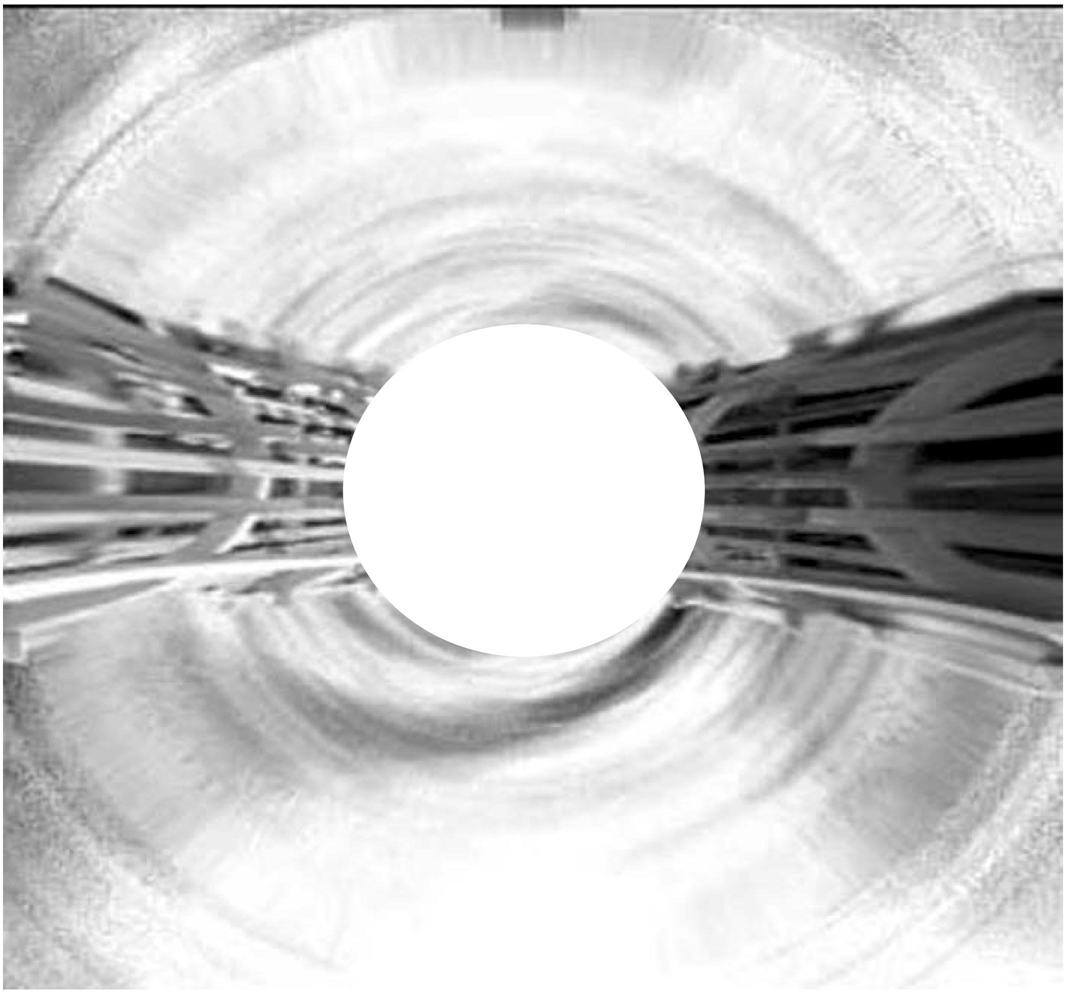
Submitted for examination November, 2001.
Corrections May, 2002.

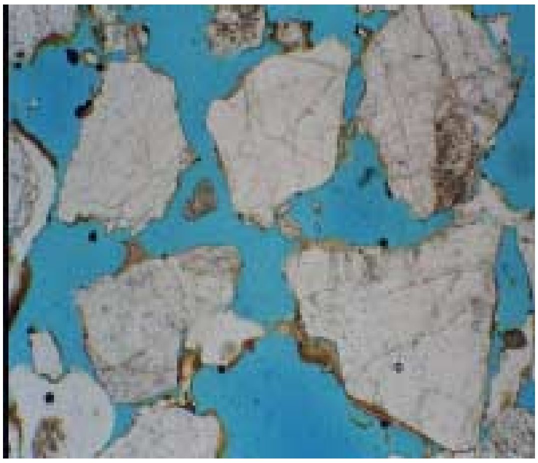
B.Sc. 1995 - University of Newcastle
B.Sc. (Honours I) 1996 - Australian National University

wendy.timms@wrl.unsw.edu.au
www.wrl.unsw.edu.au/∼wendy
UNSW Groundwater Centre, School of Civil and Environmental Engineering,
Water Research Laboratory, King St., Manly Vale, NSW, 2093, AUSTRALIA.

We know more about celestial bodies
than about the earth underfoot.

- Leonardo da Vinci d. 1519





Photomicrograph of a sandy clay aquitard.

(details in Chapter 5)

Abstract

Variable groundwater quality in complex aquifer-aquitard systems presents a challenge for sustainable groundwater development. In the Lower Murrumbidgee alluvial fan of the Murray-Darling Basin in semi arid inland Australia, shallow groundwater is saline (12 000 μS/cm) and locally contaminated by nitrate. Deep fresh aquifers (150 μS/cm), developed as an irrigation water supply, were thought to be protected from downwards leakage by laterally extensive aquitards.

However, hydrochemical sampling, augmented by historic data, revealed that aquifer salinisation (400 to 4 000 μS/cm) had occurred at some sites to 50 m depth since the mid 1980s. Aquitard windows, landscape depositional features at a scale of 10s to 100s of metres which are rarely detected by conventional investigations, were proposed as conduits for rapid downwards leakage in stressed systems.

Intensive research was conducted at the Tubbo site where downhole geophysical logging and minimally disturbed cores were used to describe a saline clayey silt to 15 m depth, an indurated clayey sand and 2 deep deposits of hard clayey silt. Fracturing was inferred by the scale dependency of aquitard permeability (K_v 10^{-11}–10^{-6} m/s). Lithological variation near the surface was delineated by electrical imaging which revealed a 40 m wide aquitard window beneath a veneer of smectite clay. Intensive monitoring of groundwater pressures in six piezometers (23–96 m depth) near the Tubbo irrigation bore and two other peizometers upgradient, indicated that the indurated clayey sand formed an effective hydraulic barrier but the deep silty deposits were spatially discontinuous.

Groundwater samples were collected before, three times during, and after the 1998–99 irrigation season. A large, but delayed TDS increase occurred in the shallow aquifer and small pulses of saline water were sustained in the middle aquifer but shortlived in the deep aquifer. Hydrochemical and isotopic data ($\delta^{13}C$, $\delta^{2}H$, $\delta^{18}O$, ^{14}C and ^{3}H) showed the middle aquifer mixing with the deep aquifer, though retaining the signature of a palaeowater. Hydrochemical changes were accounted for with PHREEQC inverse mass balance models for the shallow aquifer. Mixing of aquifer water with 20–70% saline porewater from the upper aquitard occurred, together with ion exchange and NaCl dissolution.

Based on an axisymmetric radial FEFLOW model, 5–30% of the volume pumped was accounted for by vertical leakage from the middle aquifer. Leakage from the shallow aquifer was small but significant, as it allowed high salinity water to migrate. Permeability and compressible storage measurements (S_S 10^{-5}–10^{-4} /m) were used to constrain model calibration, and to show that direct mixing occurred mainly via aquitard windows at depth, and between the shallow and middle aquifers via leaky boreholes. Fracture flow and aquifer-aquitard interaction by diffusion were of secondary importance.

Acknowledgements

Dr Ian Acworth, as supervisor played a primary role in the inception and execution of this project. His enthusiasm for field investigation and guidance in developing substance and style for publication, were invaluable and I thank him. Dr Jerzy Jankowski, co-supervisor, provided essential input throughout the project, which was greatly appreciated. His ability to see a hydrochemical story in tables of data is legendary, and taught me much.

Thanks also to our collaborators in the NSW Department of Land and Water Conservation who posed questions and provided logistical support, especially Scott Lawson, Liz Webb and Prem Kumar. Land and Water Australia are acknowledged as the main sponsor for my endeavours over 3.5 years, thanks to an Irrigation Postgraduate Scholarship. Additional funding was provided courtesy of the Natural Heritage Trust, the Department of Land and Water Conservation and the UNSW School of Civil and Environmental Engineering.

This project has been a team effort. I will always be indebted to Rolf Beck (especially for his free wisdom), Mark Groskops, John Hart, John Baird, Margaret Titterton and others at the UNSW Water Research Laboratory. This work depended on the technical expertise of others at UNSW including Dr Garreth Swarbick and Lindsay O'Keefe, Dorothy Yu, Dr Ervin Slansky and Rad Flossman, whom I thank. The support and assistance of Dr Annita Andrews and Brad McDonald at CSIRO Centre for Isotope Studies, and Mark Howden at Manly Hydraulics Laboratory was appreciated. Experience is knowledge, or so I found, thanks to discussions with Dr Neil Lavitt (formerly UNSW Groundwater Centre), Jane Coram (Federal Bureau of Rural Sciences), Dr Pauline English (CSIRO Land & Water) and Dr Noel Merrick (UTS National Centre for Groundwater Management). I am grateful to those that took the time to provide me with data and information: Dr Jeff Hanor, Dr Claude Taylor, John Thompson, Don Woolley, Genaro Piscapo, Dr Baden Williams, Andrew Love and Bill Trewhella. The willingness of local landholders to share their knowledge (and free 20 L containers of bore water at a time when Sydney's water was suspect) was very much appreciated, particularly John Gaston at Tubbo and John Lot. For my fellow research students (in order of appearance): Karen Maher, Paul Chou, Roy Soriano, Greg Dacey, Wendy McLean, James Smith, Justin Bell, Ian Brandes - many thanks for dinners out, gossip and perspective. Thanks to those who slaved with me in the field - Dr Brian Timms, Tanya Friedonhoff, Greg Dacey, and John DeMartin.

I am deeply thankful to Mum and Dad for instilling in me their love for creation. Mum has listened to my gripes and proof-read drafts, while Dad has taught me the ways of science since before I can remember.

Marge nurtured me while this thesis was born. Thankyou.

For the earth.

List of Abbreviations

AHD - Australian height datum
AMS - atomic mass spectrometer
BDL - below detection limit
BMO - best management option
BP - before present
CEC - cation ion exchange capacity
CIA - Coleambally Irrigation Area, CIC - Coleambally Irrigation Corporation
COD - chemical oxygen demand
cps - counts per second
DLWC - Department of Land and Water Conservation (formerly DWR, WRC)
DO - dissolved oxygen
ECa - apparent electrical conductivity
Eh - redox potential
EI - electrical image
BCL - below casing level
LGM - Last Glacial Maxima
LMWL - Local Meteoric Water Line
GMWL - Global Meteoric Water Line
Ma BP - million years before present
MDB - Murray Darling Basin, MDBC - Murray Darling Basin Commission
MGMC - Murrumbidgee Groundwater Management Committee
MIA - Murrumbidgee irrigation area
N - number
pMC - percent modern carbon
RMS - root mean square
SAR - sodium absorption ratio
SI - saturation indice
SWL - standing water level
SY - sustainable yield
TDS - total dissolved solids
TIC - total inorganic carbon
TTL - thermoluminescent dating
TOC - total organic carbon
TU - tritium unit (or TR - tritium ratio)
UEM - unidentified end-member
UNSW - University of New South Wales
USGS - United States Geological Survey
VSMOW - Vienna standard mean ocean water
WRL - Water Research Laboratory
WUE - water use efficiency
XRD - xray diffraction

Nomenclature

(Including dimensionality where appropriate)

ϕ - porosity (L^3/L^3)
a_v - coefficient of compressibility (LT^2/M)
c_v - coefficient of consolidation (L^2/T)
D' - aquitard thickness (L)
dh/dL - hydraulic gradient (L/L)
e_0 - void ratio (L^3/L^3)
EC - electrical conductivity
EC_a - apparent electrical conductivity
ESR = $\frac{Na}{(Ca+Mg)}$ (exchangeable sodium ratio)
h_{diff} - hydraulic diffusivity (L^2/T)
K - hydraulic conductivity (L/T)
K_v - vertical hydraulic conductivity (L/T)
K_h - horizontal hydraulic conductivity (L/T)
K_{sp} - solubility product
L - leakage factor (L)
LF - leaching factor (L/L)
MH = $\frac{Mg}{(Ca+Mg)} \times 100$ (magnesium hazard)
m_v - coefficient of volume change (LT^2/M)
P_{ex} - pore water pressure in excess of hydrostatic (M/LT^2)
Q - volumetric flux (L^3)
S - storativity
SAR = $\frac{Na}{\sqrt{(Ca+Mg)/2}}$ (sodium absorption ratio)
Ss - specific storage (1/L)
SS - soluble salts
Sy - specific yield
T - transmissivity (L^2/T)
V_a - Darcy velocity (apparent) (L/T)
V - linear seepage velocity (deep drainage) (L/T)

Contents

ABSTRACT . . . i
Acknowledgements . . . ii
List of Abbreviations . . . iii
Nomenclature . . . iv

1 INTRODUCTION 1
1.1 Knowledge gaps . . . 2
1.1.1 Leakage pathways through aquitards . . . 3
1.1.2 Identifying causes of groundwater quality variability . . . 4
1.2 Hypotheses and conceptual model . . . 5
1.3 Project objectives . . . 7
1.4 Project scope . . . 8
1.5 Overview of the Lower Murrumbidgee area . . . 9
1.5.1 Landuse and groundwater use history . . . 9
1.5.2 Challenges for groundwater management . . . 12
1.6 The Tubbo site . . . 13
1.7 Sketch of this thesis . . . 13

2 MURRUMBIDGEE GROUNDWATER ENVIRONMENT 19
2.1 Climate . . . 19
2.2 Hydrology . . . 20
2.2.1 Modern surface drainage . . . 20
2.2.2 Palaeodrainage . . . 21
2.3 Geological setting . . . 22
2.3.1 Basement geology . . . 22
2.3.2 A brief history of the Murray Basin . . . 23
2.4 Aquifer systems . . . 23
2.4.1 Renmark Formation . . . 26
2.4.2 Calivil Formation . . . 26
2.4.3 Shepparton Formation . . . 27
2.5 Aquitard deposits . . . 27
2.5.1 Mid-Tertiary permeability barriers . . . 28
2.5.2 Soils . . . 28
2.5.3 Aeolian contributions . . . 30
2.5.4 Soil permeability . . . 31
2.6 Groundwater resources . . . 32

2.6.1 Groundwater levels related to usage and allocation 33
2.6.2 Recharge sources 34
2.6.3 Implications of ^{14}C dating for recharge 37
2.7 Groundwater quality 38
2.7.1 Review of hydrochemical studies 38
2.7.2 Groundwater vulnerability 39
2.7.3 Spatial variability of saline groundwater 39
2.7.4 Observed and predicted water quality changes 40
2.8 Shallow saline groundwater 42
2.8.1 History and extent of problem 42
2.8.2 Mitigation by groundwater pumping 44
2.9 Summary 48

3 LITERATURE REVIEW 49
3.1 The nature of aquitards 49
3.1.1 Characteristics of clay minerals 50
3.1.2 Permeability 52
3.1.3 Specific storage 53
3.1.4 Pore pressure response 53
3.2 Leakage processes 55
3.2.1 Aquifer-aquitard permebility contrasts 55
3.2.2 Macropore flow 56
3.2.3 Leakage through aquitard windows 58
3.3 Hydrogeochemical role of aquitards 60
3.3.1 Matrix interactions 60
3.3.2 Solute transport processes 61
3.3.3 Effects on adjacent aquifers 64
3.4 Accounting for groundwater quality variability 65
3.4.1 Trend analysis 65
3.4.2 Well installation trauma 67
3.4.3 Groundwater vs. well water quality 67
3.4.4 Case studies of temporal changes 69
3.5 Causes of groundwater quality variability 72
3.5.1 Episodic recharge 73
3.5.2 Agricultural development 73
3.5.3 Groundwater flow enhanced by pumping 75
3.5.4 Water quality improvement after pumping ceases 76
3.6 Towards an integrated approach 78
3.6.1 Testing flow models with hydrochemical information 79
3.6.2 Groundwater pressure vs. quality equilbrium. 80
3.7 Summary 81

4 FIELD AND LABORATORY TECHNIQUES 82
4.1 Hydrogeology 82
4.1.1 Geophysical investigation 82
4.1.2 Drilling, coring and piezometer installation 83

4.1.3 Pump testing and permeability analysis . . . 85
4.1.4 Solid phase analysis . . . 89
4.2 Hydrochemistry . . . 90
4.2.1 Sampling procedures . . . 90
4.2.2 Laboratory analysis . . . 93
4.3 Isotope techniques . . . 94
4.3.1 Carbon-14 dating of groundwater . . . 97
4.3.2 Tritium dating of groundwater . . . 98
4.4 Quality assurance . . . 99

5 AQUITARD HYDROGEOLOGY 100
5.1 Results . . . 100
5.1.1 Borehole geophysics . . . 100
5.1.2 Electrical images . . . 104
5.1.3 Lithological sections . . . 105
5.1.4 Physical description . . . 106
5.1.5 Thin section analysis of indurated clayey sand . . . 108
5.1.6 Falling head permeameter testing of core samples . . . 109
5.1.7 Oedometer testing of core samples . . . 111
5.1.8 Specific yield and effective porosity . . . 113
5.1.9 Slug test of lower silt unit . . . 116
5.2 Aquitard permeability . . . 116
5.2.1 Evidence for leakage through aquitard matrix. . . . 116
5.2.2 Related to consolidation state . . . 117
5.2.3 Scale dependency . . . 118
5.3 Aquitard windows . . . 119
5.3.1 Extent of the indurated clayey sand . . . 120
5.4 Stratigraphic nomenclature . . . 121
5.5 Summary . . . 122

6 GROUNDWATER HYDRODYNAMICS 123
6.1 Results . . . 123
6.1.1 Regional groundwater hydrographs . . . 123
6.1.2 Pumping history at Tubbo . . . 125
6.1.3 Groundwater levels . . . 126
6.1.4 Gradients . . . 129
6.1.5 Drawdown analysis . . . 130
6.1.6 Pump test analysis . . . 133
6.1.7 Slug tests of aquifers . . . 134
6.2 Evidence for window leakage . . . 135
6.3 Profile and radial flow modelling . . . 135
6.3.1 Model purpose . . . 136
6.3.2 Conceptual model . . . 137
6.3.3 Model development . . . 138
6.3.4 Calibration . . . 145
6.4 Modelled water balance . . . 146

6.4.1 Sensitivity to aquitard permeability and compressibility 149
6.4.2 Sensitivity to aquitard window width 149
6.4.3 Sensitivity to bore casing leakage 153
6.5 Model limitations 153
6.6 Summary 155

7 HYDROCHEMICAL VARIABILITY 156
7.1 Results 156
7.1.1 Unstable parameters 156
7.1.2 Major ions 158
7.1.3 Minor and trace elements 168
7.1.4 Stable isotopes 171
7.1.5 Radioisotopes 175
7.2 Accounting for hydrochemical variability 177
7.2.1 Artifacts of sampling and well installation trauma 177
7.2.2 Short term trends 180
7.2.3 Medium term trends 182
7.3 Geochemical processes 183
7.3.1 Leakage from clayey aquitards 183
7.3.2 Mixing and dilution 183
7.3.3 Dissolution and precipitation 186
7.3.4 Ion exchange 186
7.4 Mass balance modelling 188
7.4.1 Quantifying mixing and matrix interaction 188
7.4.2 Assumptions 189
7.4.3 Rice Mill site 190
7.4.4 Tubbo site 192
7.5 Summary 193

8 SYNTHESIS 194
8.1 Solute transport processes 194
8.1.1 Flux through the vadose zone to the shallow aquifer 194
8.1.2 Advective flux between the shallow and middle aquifers 196
8.1.3 Leakage through aquitard windows between the middle and deep aquifers 197
8.1.4 Diffusive flux between aquitards and deep aquifers 197
8.2 Groundwater water quality degradation related to stress 200
8.3 Hydrochemical verification of flow modelling 203
8.4 Summary 204

9 CONCLUSION 205
9.1 Key findings 205
9.2 Verification of hypotheses and revised conceptual model 207
9.3 Lessons learnt from this research 209
9.4 Future challenges 210
9.5 Conclusion 211

Personal communications 212

Bibliography 212

A Site summaries of key geophysical and hydrochemical data 234

B Bore and piezometer details 235

C Enclosed CD-ROM "Publications & Datasets" 238
C.1 Publications and reporting 238
C.2 Hydrogeology 239
C.3 Hydrochemistry 239
C.4 Lab reports 239

List of Figures

1.1 Working conceptual models of solute transport in aquifer-aquitard systems depicting the relative importance of vertical leakage and diffusive versus advective solute transport under natural conditions. 6
1.2 Location maps for the study area and nested piezometer sites 10
1.3 Aerial photograph of the Tubbo site and surrounding features, including locations of piezometer installations and electrical image lines. 11
1.4 Aquitard and aquifer geometry and piezometer installations at the Tubbo site 14
1.5 Surface and groundwater resources supplying agriculture in the Lower Murrumbidgee area (photographs). 17
1.6 The Tubbo site (photographs) 18

2.1 Average annual rainfall at Leeton, 1914–2000. 20
2.2 Palaeodrainage and interchannel plains relative to study sites. 21
2.3 Murray Basin groundwater section. 24
2.4 Palaeofacies of the Murray Basin depicting development of alluvial fans at the margins of the Riverine Plain. 25
2.5 Comparison of estimated connate and present pore-water salinities of aquifer-aquitard sequence at Piangil West site. 30
2.6 Groundwater development in the Lower Murrumbidgee study area showing near steady rate of increase in the number of bores between 1940–2000, and increased allocations and usage since the mid 1990s. 33
2.7 Groundwater residence times for the Calivil aquifer inferred from carbon-14. 37
2.8 Spatial distribution of groundwater salinity, Shepparton Formation, September, 1999. 40
2.9 Observed changes in salinity with depth depicting downwards movement of salinity bulge between 1970 and 1980 41
2.10 Predicted salinity of the deep pumped aquifers with natural leakage and full mixing of the shallow saline zone 41
2.11 Actual and projected areas of the CIA with watertable depth <2 m from surface. 44
2.12 Projected salinity increase of pumped groundwater over time. 47

3.1 Saline intrusion along preferrential flowpaths in a multi-layered aquifer, Atlantic coast, US. 77

4.1 HQ triple barrel system for recovery of minimally disturbed cores. 84

4.2 Groundwater field methods (photographs). 86
4.3 Permeability testing methods (photographs). 87

5.1 Borehole geophysical logs from selected sites in the alluvial fan showing regional variability. 101
5.2 Geophysical logs and physical parameters derived from core analysis for UNSW-B6. 103
5.3 Electrical image for Rice Mill site. 104
5.4 Electrical images (inverse models only) showing palaeodrainage features in upper silt unit. 105
5.5 Photographs of heterogeneities in silt core. 107
5.6 Thin-section microphotographs of indurated clayey sand with pore spaces stained blue. 108
5.7 Example of permeability variation due to gradient and temperature changes during a falling head permeameter test. 110
5.8 Pore pressure variation with depth at Tubbo site, with calculated total and effective stresses. 112
5.9 Example of consolidation test where settlement occurs due to dewatering of the aquitard matrix with increasing normal stress. 112
5.10 Specific yield estimate based on texture. 115
5.11 Scale dependent hydraulic conductivity for the indurated clayey sand, Lower Shepparton and Calivil aquitards. 118

6.1 Hydrographs for selected nested piezometer sites, 1971–2000. 124
6.2 Groundwater usage by the Tubbo irrigation bore GW59113, 1990–1998. . . 126
6.3 Hydrographs for nested piezometers UNSW-B1–B6, Tubbo site. 127
6.4 Hydrographs for the DLWC monitoring bores 30282, 30284, Tubbo site. . . 128
6.5 Head differences and gradients during the 1998–1999 irrigation season, Tubbo site. See Figure 6.3 for pumping history. 130
6.6 Drawdown during pump testing at Tubbo site, 11–24 September 1998. . . . 132
6.7 Radial axisymmetric flow modelling. 136
6.8 Setup of layers and boundary conditions for FEFLOW modelling at Tubbo site. 138
6.9 Observed and modelled hydraulic heads at Tubbo during the 1st pumping cycle, 11–26th September, 1998. 139
6.10 Example of axisymmetric FEFLOW model for a multi-layered system with an aquitard window, showing hydraulic response to pumping stress. 141
6.11 Mesh setup showing radially expanding elements and mesh enrichment near the pumping bore. 142
6.12 Water balance for 1st pumping cycle, with no aquitard windows, 11–26th September, 1998. 147
6.13 Water balance for 1st pumping cycle, with aquitard windows, 11–26th September, 1998. The range of flux specified for each flowpath is a summary of numerous individual models. 148
6.14 Sensitivity of aquitard drawdown to compressible storage with no aquitard windows (K = 10^{-7} m/s). 150

6.15 Sensitivity of aquitard drawdown to permeability including aquitard windows (Ss = 3x10^{-4}). 150
6.16 Vertical flux related to aquitard window width, showing narrow windows greatly increase vertical flux and a threashold width beyond which little increase in vertical flux occurs. 152
6.17 Cross-sections through layer 6 aquitard, showing high sensitivity of groundwater velocity, and insensitivity of hydraulic head to a 70m wide window. . 152

7.1 Spatial variability of salinity and watertype, February 1999 & 2001. 157
7.2 Variability of unstable parameters with depth at Tubbo site, 1998/99. . . . 159
7.3 Salinity variability in shallow and deep aquifers at selected sites, 1970–1999 160
7.4 Piper diagrams showing hydrochemical variation at selected sites 1970–2001 162
7.5 Major ion variability at Rice Mill site, 1980–1999. 163
7.6 Depth profiles of major species at Tubbo site, 1998–1999. 164
7.7 Major ion variability at Tubbo site, Sep-98 to May-99. 166
7.8 Trace element variability with depth at Tubbo site, Sep-98 to May-99. . . . 170
7.9 Oxygen-18 and deuterium data from Lower Murrumbidgee. 172
7.10 Oxygen-18 versus chloride for Murrumbidgee River and groundwaters at Gundaline and Tubbo sites, Sep-98, Oct-98 and Feb-99. 174
7.11 Variability of δ^{13}C, ^{14}C and tritium isotopes with depth, Tubbo site, Oct-98 and Feb-99. 176
7.12 Groundwater EC during bore purging prior to sampling, February, 1999. . . 179
7.13 Piper diagram of short term hydrochemical changes at Tubbo site, Sep-98 to May-99. 181
7.14 Mixing diagram for low chloride river and deep aquifer water with high chloride shallow groundwater. 184
7.15 Saturation indices for groundwater and surface waters, Feb-98. 187
7.16 Ionic ratio bivariate plots for ground and surface waters, Feb-99 & Feb-01. . 187

8.1 Relationship betweeen major ion chemistry of porewater determined from Upper Shepparton aquitard core samples and groundwater from the underlying Upper Shepparton aquifer. 195
8.2 Range of Darcy flux for the indurated sand aquitard under natural and induced gradients, based on measured and calibrated model permeability values. 196
8.3 Relationship betweeen major ion chemistry of porewater determined from aquitard core samples and groundwater from piezometers within the Calivil aquitard and adjacent aquifers. 199
8.4 Correlation of groundwater level and salinity trends. A Rice Mill site, 1979–2001 B CIA pump site, 1990–2001. 201
8.5 Shallow groundwater salinity trends compared with water salinity criteria, 1970–2001 202

9.1 Conceptual models of an aquifer-aquitard system depicting the relative importance of vertical leakage and diffusive versus advective solute transport under natural and developed conditions. 208

List of Tables

1.1 Piezometer installation details at Tubbo site. 15

2.1 Aquitards in the south-eastern Murray Darling Basin. 29
2.2 Vertical hydraulic conductivity of soil and aquitards in the Murray Basin and Lower Murrumbidgee areas. 32
2.3 Recharge sources for the Lower Murrumbidgee 36
2.4 Induced leakage at the Coleambally deep bore 36

3.1 Distinguishing aquitards and aquifers 50
3.2 Characteristics of common clay minerals. 51
3.3 Relative aquifer and aquitard K, selected examples 56
3.4 Examples of aquitard effects on regional groundwater resources 57
3.5 Importance of factors that result in under or overestimation of ^{14}C groundwater dating 61
3.6 Examples of aquitard effects on groundwater chemistry 64
3.7 Selected case studies of groundwater quality changes over time 70
3.8 Magnitude of groundwater salinity increase in selected areas 71

4.1 Details of pressure transducers deployed at Tubbo site on the 7-Aug-98. . . 85
4.2 Field electrode calibrations 91
4.3 Summary of water analysis methods 93
4.4 Potential errors in δ^{13}C sample preparation. 96
4.5 Interpretation of tritium activity in Australian rainwater assuming no mixing. 98

5.1 Basement geology for bores where data is available. 106
5.2 Bulk mineralogy of the indurated clayey sand at Tubbo 107
5.3 Clay mineralogy of bulk and clay fraction for indurated clayey sand at Tubbo 107
5.4 Summary of permeability test results for core samples. 111
5.5 Measured c_v values compared with typical values for aquitard units. 113
5.6 Measured m_v values compared with typical values for aquitard units. 114
5.7 Summary of derived hydraulic parameters for aquitard units. 114

6.1 Location of bores, Murrumbidgee River and main irrigation channel, relative to Tubbo irrigation bore GW59113. 129
6.2 Drawdown delays observed at Tubbo site after start of pumping, 11th September, 1998. 131

6.3 Hydraulic properties of aquifers in the Murray Basin and Lower Murrumbidgee areas. . . . 144
6.4 Storativity of aquifers in the Murray Basin and Lower Murrumbidgee areas. 144
6.5 Range of hydraulic properties tested during calibration of Tubbo model. . . 145
6.6 Examples of the relationship between aquitard window width and the proportion of vertical and lateral flow. . . . 151
6.7 Vertical leakage via boreholes. . . . 153

7.1 Hydrochemical data used in the study. . . . 157
7.2 Salinity trends at nested piezometer sites, pre-1980–1999. . . . 161
7.3 Initial and final watertypes for surface and groundwaters at Tubbo, Sep-98 and May-99. . . . 167
7.4 Strontium and manganese concentrations detected, Feb-99 and Feb-01. . . . 169
7.5 Copper, lithium and boron trace concentrations detected, Feb-99. . . . 170
7.6 Slope and deuterium excess for Australian rainfall isotope monitoring stations. . . . 173
7.7 Stable Isotope results for Tubbo site, Sep-98, Oct-98 and Feb-99. . . . 174
7.8 Radioisotopes at Tubbo site, Feb-99, including apparent and Pearson corrected ages. . . . 177
7.9 Indicators of bore casing corrosion and leakage, February, 1999. . . . 180
7.10 Mass balance model development for Rice Mill site, between Initial 36372-$1_{(Sep-86)}$ and Final 36372-$1_{(Feb-99)}$ Aquifer waters 191
7.11 Mass balance model development for Tubbo site, between Initials B-$1_{(Oct-98)}$, Shepparton Clay porewater and Final $B1_{May99}$ 192

B.1 Bore installation details for bores sampled. . . . 236

Chapter 1

INTRODUCTION

Water resources in Australia's inland basins are increasingly stressed by development, potentially compromising environmental flow needs, inter-generational access to potable water, and the sustainability of agriculture itself.

Sustainable irrigated agriculture was defined by Meyer (1992) as:

> "...the protection of our land and water resources whilst developing the profitability of the irrigation industry."

Sustainable irrigated agriculture was identified by the National Farmers Federation (NFF 1996) as one of Australia's key environmental issues of the 1990s, and in an international context as of critical importance in terms of water quality (OECD 1998). Demand for water has been growing at more than twice the rate of population increase during this century and most of the increase in the world's food supply has come from the expansion of irrigated lands. Protection of Australia's food producing capacity is an issue relevant to both rural and urban Australians, as well as our neighboring countries and trading partners.

One of the primary concerns of agricultural sustainability is safeguarding water quality for a range of applications. For example, the toxicity of elevated sodium, magnesium and chloride levels to specific crops is well known. Salinisation has led to the collapse of agricultural economies around the world and throughout history (Hillel 1991). Mesopotamian civilisation declined between 4 000 and 1 700 BC as salinisation affected the Tigris and Euphrates irrigation areas (Ghassemi et al. 1995). The range of water quality issues beyond salinity should not be forgotten. For instance, livestock health may be affected by excessive nitrate concentration and the contamination of potable water supplies with pesticides such as atrazine is an issue in some agricultural communities.

In Australia, the driest inhabited continent, groundwater accounts for 18% of total water use, of which half is used for irrigation purposes. Over 1 million people rely on groundwater as their principle drinking water source (Smith 1999). Although groundwater may not be immediately vulnerable as are surface waters, contaminates can permeate downwards through the unsaturated vadose zone and from shallow to deep aquifers. Groundwater quality may deteriorate to such an extent that mitigation is either uneconomic or not physically possible.

Sustainable groundwater development

Unsustainable development of groundwater resources has occured in many areas around the world. The concept of unsustainable groundwater development has not been adequately defined, but may be considered equivalent to over-draft, over-exploitation, or groundwater mining (Custodio 1991). The most recent local definition of sustainable yield, considered to be a proportion of the long term average annual recharge, (DLWC May, 2001) is:

> "...the groundwater abstraction regime, measured over a specified planning timeframe, that allows acceptable levels of stress, and protects the higher value users that have a dependency on the water."

1.1 Knowledge gaps

There is a common perception that to achieve sustainable agriculture requires socio-economic incentives to implement established best management options (BMO's) through personal, community and political willpower. Whilst all of these factors are critically important, the increasing complexity of resource management problems inevitably means that our existing knowledge and technical skills base is far from adequate. The level of investment in monitoring and research should reflect the millions of dollars at stake in agriculture so that sound science underpins the drive toward sustainability.

There is great uncertainty for instance, about the volume of water which leaks between aquifer systems, in quantifying surface-groundwater interactions and in defining the sustainable yield (SY) of a groundwater system in terms of both groundwater quantity and quality. This research targets gaps in fundamental knowledge that were in part identified by local resource managers and users in the Lower Murrumbidgee area of Australia's Murray-Darling Basin, and which in many respects, are common with aquifer-aquitard systems in Portugal, the Mexico City basin, the Canadian prairies and elsewhere.

1.1.1 Leakage pathways through aquitards

Historically, groundwater investigations have focused on high yielding aquifers rather than the aquitards (confining units of low permeability) that control both the quantity and quality of recharge. The recognition of the importance of the role of aquitards is exemplified by the recent theme issue of Hydrogeology Journal on confining units (Remenda 2001).

One of the first comprehensive accounts of the permeability of clays and shale deposits where flux occurred through the matrix pore space, was presented by Neuzil (1986). Other research has focused on hydraulic responses within aquitards to loads like changing groundwater pressure in surrounding aquifers (eg. van der Kamp & Maathuis 1991, Rudolph & Frind 1991). Solute transport through aquitards, dominated by matrix diffusion, has been studied in the context of clay-lined waste disposal (eg. Johnson et al. 1989, Rowe et al. 1995) and over geologic time through thick clay aquitards (eg. Hendry & Wassenaar 2000).

A relatively homogeneous groundwater flow system, with regionally extensive aquitard units is often assumed in groundwater contaminant investigations, particularly on a regional scale. However, spatially variable permeability is one of the greatest uncertainties in contaminant transport (Smith & Schwartz 1981), and it has been recognised that heterogeneity becomes increasing important as the scale of investigation decreases (eg. Back 1986). The impact of heterogeneity on flow paths is likely to be most evident where flow occurs over a relatively short period of time at site scale or local scale (Back et al. 1993).

Aquitard fractures, a microscopic form of heterogeneity, have recieved much attention (eg. Grisak & Cherry 1975, Hendry 1982, Rudolph et al. 1991, Keller et al. 1991, Harrison et al. 1992). Laboratory testing of representative fractured clay samples and field pit characterisation of fracture density, together with monitoring of groundwater quality have shown that even non-visible fractures are able to rapidly transmit significant quantities of contaminates.

The relative importance of larger scale heterogeneity due to sedimentary structures, however, has not been adequately assessed (Gerber et al. 2001). Some studies have considered the role of sedimentary structures in glacial till, of <1 mm to $>10^3$ m scale (eg. Boyce & Eyles 2000). But leakage through 'landscape holes' in other depositional environments has been neglected (Williams et al. 2001). 'Holes' like aquitard windows, which may be 1–100 m scale, are not possible to study in the laboratory and are inherently difficult to investigate in the field. An integrated hydrogeologic appraisal which reconstructs geologic facies and assigns appropriate field permeability values may be possible (eg. Fogg 1986), or for very small and shallow groundwater systems, numerous closely spaced permeability measurements may be practical to characterise heterogeneity (eg. Jankowksi & Beck 2000).

Numerical models for the complex Waterloo moraine in Canada demonstrated well capture zones to be highly sensitive to the presence or absence of aquitard windows, defined by comprehensive borehole data analysis (Martin & Frind 1998). If sufficient data are not available however, the existence of aquitard windows and the inadequacy of a simple conceptual model may not become apparent until after groundwater contamination has occurred (eg. MSU 2000).

Trewhella (1989) pointed out that little was known about Late Quaternary aquitards which overlie important irrigation aquifers of the eastern Murray-Darling Basin.

> *"Little is known about the aquitards between the aquifers... there is (sic) virtually no data about aquitard salinity at depth...this may prove to be an important data gap if development of the underlying deep aquifers is encouraged, and significant leakage is actually transmitted through the aquitards."*

Until recently, only two studies on the Australian continent had focused on aquitards important for water resources.The first study to associate changing groundwater quality with leakage from aquitards was Stadter & Love (1989) who documented increased aquifer salinity in the Tatiara irrigation area of South Australia. Jones et al. (1994) identified the Geera clay, in the central Murray-Darling Basin as an important salt sink for the overlying Parilla sand aquifer, and salt source for confined aquifers.

Intensive study of aquitards is currently in progress in three Australian irrigation areas: the Otway Basin of south-east South Australia (Love et al. 1995; Brown et al. 2001; Harrington et al. 2001), and in catchments of the Murray-Darling Basin including the Upper Namoi (Acworth & Jankowski 1997; Timms 1998; Acworth & Beasley 1998) and the Lower Murrumbidgee.

1.1.2 Identifying causes of groundwater quality variability

Groundwater quality variability is generally described by snapshot studies that sample a large number of bores scattered over a wide area on a limited number of occassions (often once only). Such an approach is a useful step in identifying sites where groundwater quality parameters exceed management guidelines (Loftis 1996), but more intensive study is required to establish a cause-effect relationship (Iman & Conover 1983) and to identify the relative importance of factors which contribute to changing groundwater quality over time. Due to logistical challenges, few groundwater quality studies have achieved this, either over the medium term (eg. 8 years of monthly data from a New Zealand irrigation area by Close 1987), or over the short term (eg. continuous sample collection for several hours near an irrigation bore in Nebraska by Chen et al. 1998).

A number of snapshot groundwater quality studies of aquifers in the Murray-Darling Basin have documented the spatial distribution of contaminants such as salt, nitrate and atrazine (eg. Bauld et al. 1995, Jiwan & Gates 1995, Schmidt et al. 1996, Timms 1997). However, co-ordinated and on going groundwater quality monitoring programs are lacking, despite attempts to establish national frameworks (eg. Keary 1993). Evans & Bauld (1993) recommended that a groundwater quality assessment program should include: priority for vulnerable aquifer systems, measurement of the widest possible range of water quality parameters and maintenance of a nationally consistent integrated database.

1.2 Hypotheses and conceptual model

A hypothesis and working conceptual model were proposed to describe vertical leakage processes in a complex aquifer-aquitard system. As Figure 1.1 depicts, vertical leakage is generally considered to be a minor component of the water balance, relative to surface fluxes and lateral flow in aquifers. Natural downward flux through a continuous aquitard layer are small and likely to be dominated by diffusive solute transport. However, a fractured or discontinuous aquitard may not be apparent under naturally low vertical leakage gradients. Vertical stress induced by pumping from the deep aquifer is therefore required to distinguish between the three possible aquitard models A1, B1 and C1.

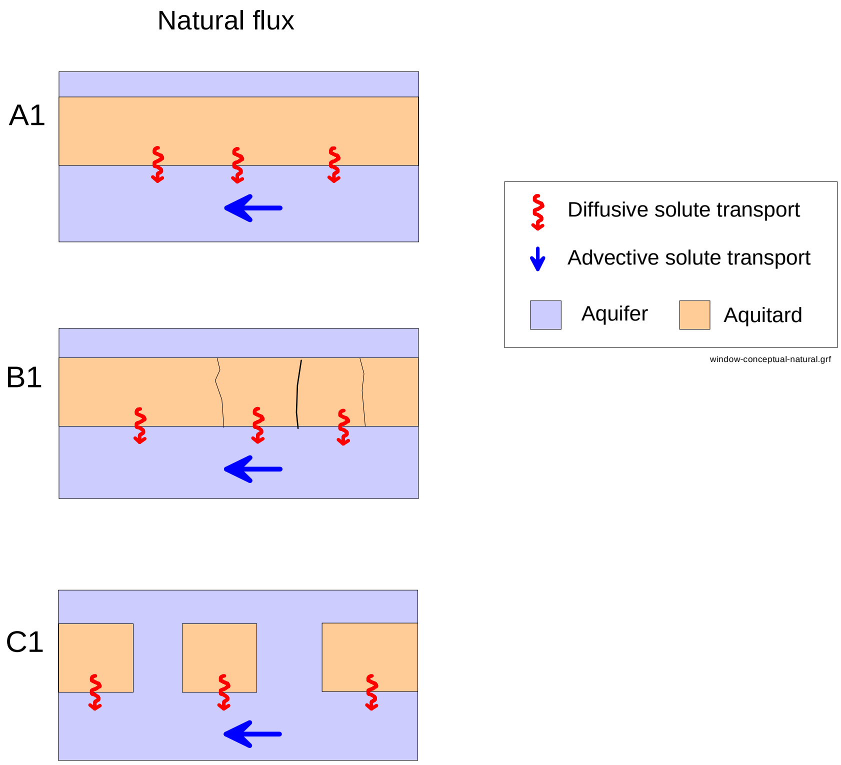


Figure 1.1: Working conceptual model of solute transport in layered aquifer-aquitard systems depicting the relative importance of vertical leakage and diffusive versus advective solute transport under natural conditions. (Flux is proportional to the number and size of arrows). **A** Continuous aquitard layer **B** Fractured aquitard. **C** Discontinuous aquitard with windows.

On the basis of these three possible conceptual models for aquitard-aquifer systems in alluvial deposits, the following hypotheses were proposed.

1. *Aquitard windows and other rapid flow pathways only become evident by applying vertical stress within a groundwater system.*
2. *Significant vertical flux of contaminants may occur when a heterogeneous aquifer-aquitard system is stressed and where groundwater quality varies with depth.*

Landscape depositional features, such as aquitards occur at a scale of 10s to 100s of metres, are rarely detected by conventional investigations. However, aquitard windows may provide a conduit for rapid downwards leakage in stressed groundwater systems.

1.3 Project objectives

The primary objective of this work was to test these hypotheses by application of field, laboratory and numerical modelling techniques. A test site at Tubbo was located in the Lower Murrumbidgee alluvial aquifer-aquitard system of the Murray-Darling Basin (Figures 1.2 and 1.3), as described in Sections 1.5 and 1.6.

Key project objectives were to:

- Characterise shallow aquitard deposits and document direct evidence for aquitard windows from surface and downhole geophysical techniques and analysis of core obtained during drilling.
- Investigate permeability contrasts between aquifer and aquitard deposits at a number of scales.
- Describe groundwater dynamics within a complex aquifer-aquitard system by installation of nested piezometers to monitor the impact of high volume groundwater abstraction.
- Quantify pore pressure changes throughout the system with analytical and numerical modelling tools.
- Detect vertical leakage through aquitard windows during periods of groundwater stress by monitoring hydrochemical changes within and between specific aquifers and aquitards at several stages before, during and after an irrigation season.
- Account for hydrochemical changes by applying mass balance geochemical models, verified with independent environmental isotopic data.

Whilst the primary objectives of the project were to investigate fundamental groundwater processes, the implications for sustainable groundwater management were also considered. For example, it was anticipated that examining the relationship between groundwater quantity and quality could lead to improved guidelines for the maintenance of beneficial use in a stressed groundwater system. Furthermore, management actions could be prioritized on the basis of key risk factors identified for groundwater quality decline under natural and developed conditions. Finally, it was recognised that this research would identify readily transferable principles and technical methodologies applicable to similar aquifer-aquitard environments within and beyond the Murray-Darling Basin.

Preliminary investigation for this project documented water quality parameters for specific layers of the aquifer-aquitard system (Timms et al. 1999) and identified the upper aquitard as the primary salt source contributing to short and long term groundwater quality deterioration (Timms et al. 2000). The development of a local flow model constrained by isotopic evidence, was introduced by Timms et al. (2001) and has been further refined here. The model was used to evaluate the relative importance of induced flux through aquitard windows and leaky bore casings.

1.4 Project scope

The scope of the project was focused at spatial and temporal scales that were suited to evaluating this proposal.

An alluvial fan deposit, characterised by complex stratigraphic architecture was the most suitable environment in which to investigate the significance of landscape heterogeneity in a groundwater system. The Lower Murrumbidgee alluvial fan proved to be ideal for detecting leakage through aquitard windows, given that groundwater quality varied with depth and the system has been increasingly stressed by groundwater development in recent years. The investigation focused on semi-consolidated fine grained deposits of Pleistocene-Pliocene age between the surface and 85 m depth.

A regional and medium term context for the project was established along a 30 km wide cross section and a 120 km longitudinal section from the near the apex to the toe of the alluvial fan (Figure 1.2). The status of groundwater resources in this area was assessed from available hydrographic and hydrochemical data during the past 30 years. This information was augmented by downhole and electrical image geophysical surveys at selected sites, and a 'snap-shot' hydrochemical sampling of 35 observation and irrigation bores, during a high stress period, February 1999.

Intensive field work was focused on the Tubbo site, located downstream of the fan apex (Figures 1.2, 1.3). Here, lithological heterogeneity was investigated at a scale of 10s to 100s of metres. Much of the site investigation was completed during a 12 month period, with the irrigation bore intermittently active between September 1998 and March 1999. A suite of appropriate inorganic chemical species and stable and radioactive environmental isotopes were selected to characterise hydrochemical variability with depth and over time. A comprehensive dataset was collected that enabled numerical flow models and mass balance geochemical models to be developed that quantified vertical leakage and mixing.

1.5 Overview of the Lower Murrumbidgee area

The Murrumbidgee catchment is a major drainage division of the Murray-Darling Basin in the semi-arid inland of south-east Australia (Figure 1.2). The Lower Murrumbidgee alluvial fan formed at the eastern margin of the Riverine Plain, under which lies the extensive Murray groundwater basin.

The Murrumbidgee catchment encompasses an area of 84 000 km^2, of which the lower catchment accounts for 50 000 km^2 (Wallbrink et al. 1998). The alluvial fan covers approximately 6 500 km^2, extending about 120 km downstream from Narrandera, near the margin of the geologic structure of the Murray Basin (WRC 1984).

The Murrumbidgee River flows for 1 600 km, from its headwaters in the Snowy Mountains, though dissected and rolling foothills, and across the Riverine Plain to its confluence with the Murray River which discharges to the Southern Ocean off South Australia. Natural flow in the Murrumbidgee River is augmented by the Snowy Mountains Hydroelectric scheme which redirects easterly flowing waters via 10 000 km of channels for western irrigation schemes including the the Coleambally Irrigation Area (CIA).

Photographs of typical irrigation features in the area are provided at the end of Chapter 1 (Figure 1.7).

1.5.1 Landuse and groundwater use history

Land use in the area includes sheep and cattle grazing, irrigated horticulture and cropping, dryland cropping, organic food growing, forestry and conservation reserves. Agricultural production in the Murrumbidgee catchment exceeds $1 billion p.a. (MCM 2001), accounting for 25% of the state of New South Wales fruit and vegetable production, 42% of the state's grapes and 50% of the nation's rice.

The Lower Murrubmidgee area was originally settled in 1840 (van der Lelij 1989). Irrigation on a large scale commenced in the early 1900s with the Murrumbidgee Irrigation area near Griffith. The 79 000 ha of the CIA was the last of several large-scale government sponsored irrigation development, spawned from the Snowy Mountain Scheme (Hallows 1987). The CIA was developed between 1960 and 1973, with several hundred farms originally established for sheep and cereal cropping but which were largely replaced in the mid-1970s by rice watered exclusively by flood irrigation.

Groundwater was first utilised in the 1850s by way of shallow timber-shored hand dug wells for domestic and limited stock supply. By the 1960s windmill pumps were ubiqutous in pastures distant from the river (Kelly 1978; Ronald 1960). Technological limitations on

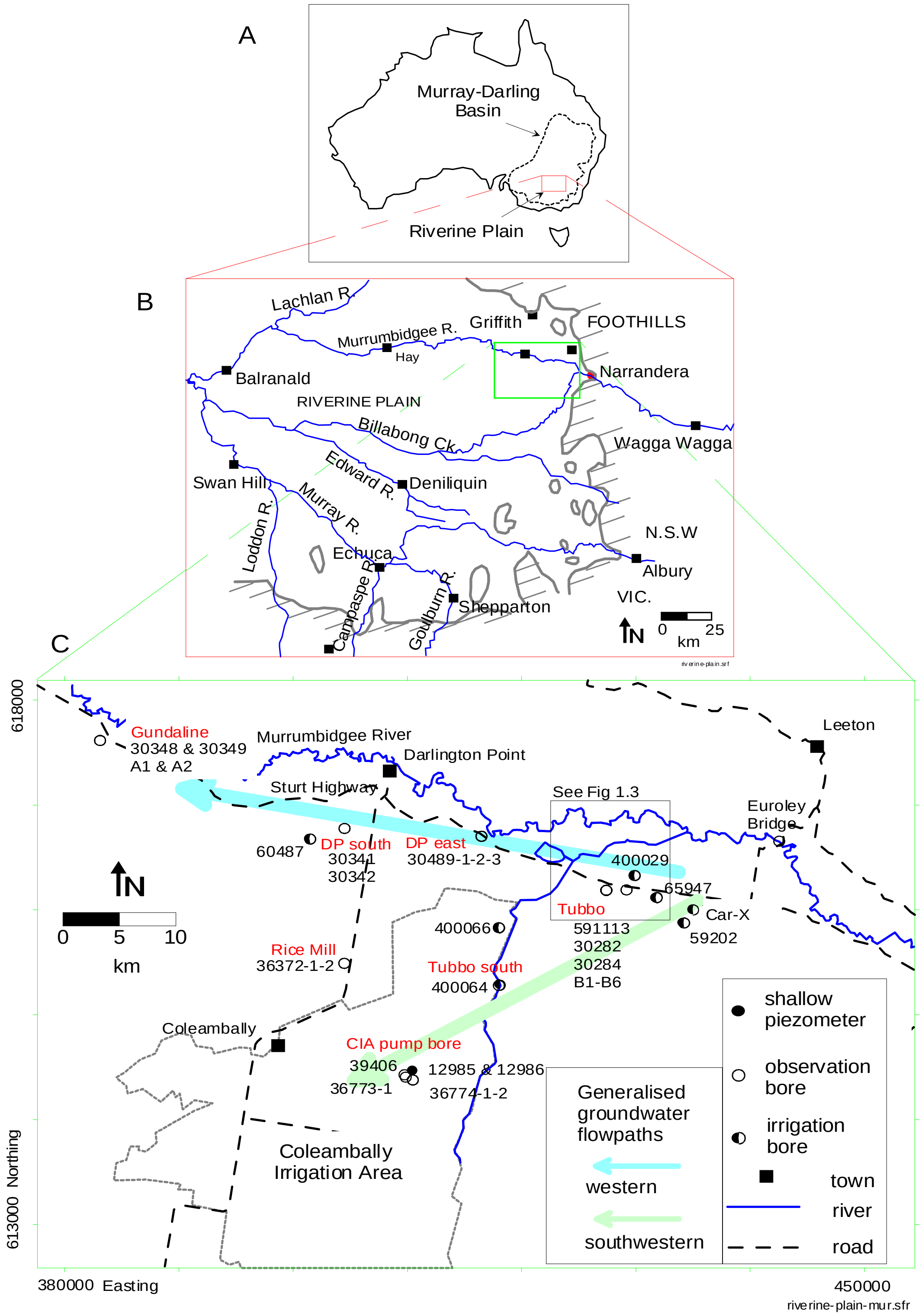


Figure 1.2: Location maps **A** The Murray-Darling Basin of Australia **B** The Lower Murrumbidgee alluvial fan at the edge of the Riverine Plain **C** Locations of observation and irrigation bores and nested piezometer sites included in the study relative to major western and southwestern flowpaths.

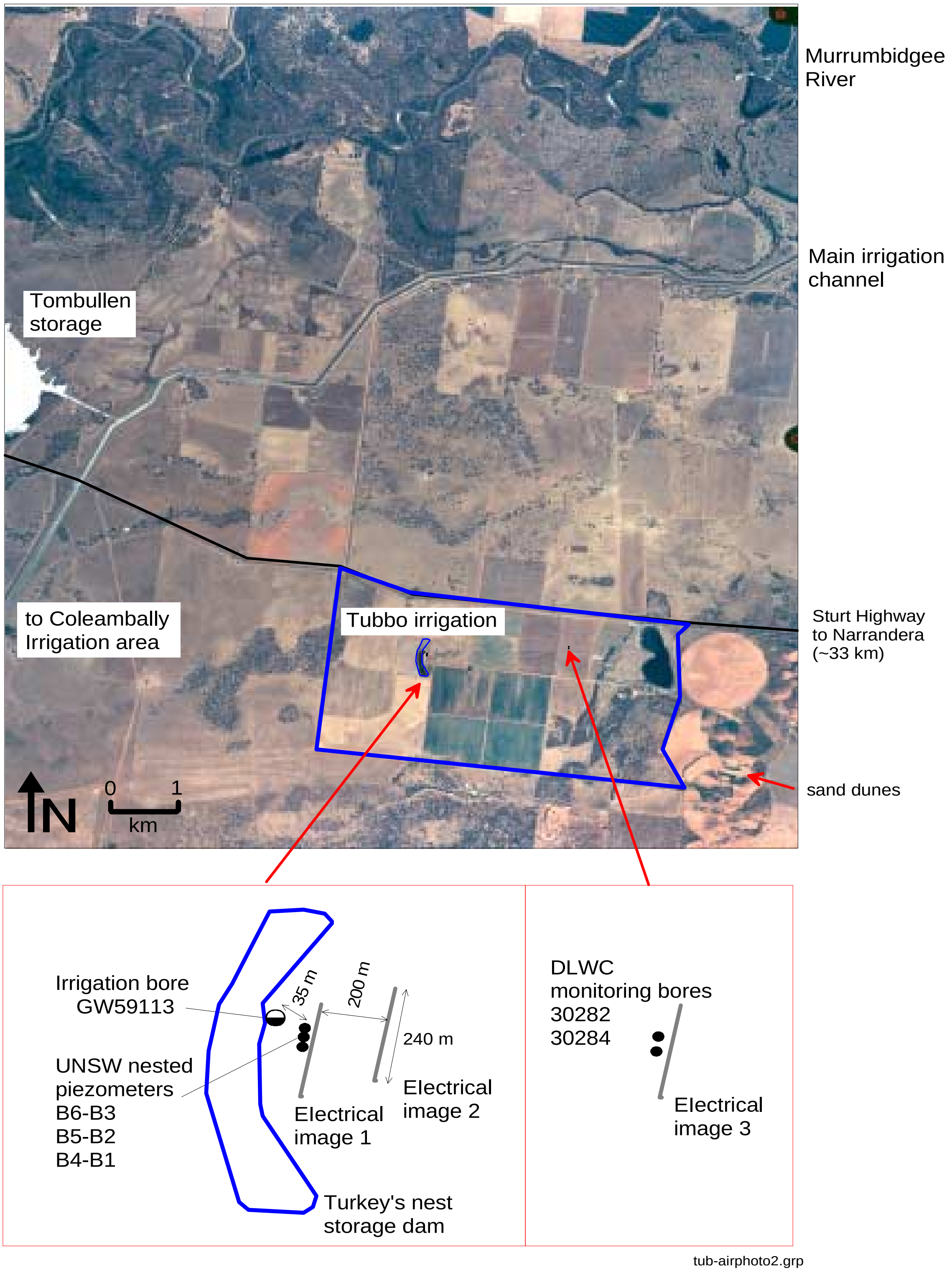


Figure 1.3: Aerial photograph of the Tubbo site and surrounding features, including locations of piezometer installations and electrical image lines. (Yanco, NSW 4176, 1:50 000 18.12.93).

drilling and pumping preserved the deep multi-layered aquifers from pumping until the 1970s. After this time groundwater was developed mainly in areas away from existing irrigation channels. Deep groundwater pumping has become an active strategy only in the last decade within the irrigation areas, where it is used for shallow watertable management, and to augment limited surface water supplies for irrigation.

1.5.2 Challenges for groundwater management

The Lower Murrumbidge area faces apparently contradictory groundwater sustainability problems with declining groundwater pressures in some areas and rising watertables elsewhere. In the CIA, watertables were at about 20 m depth when development began in the 1960s. Hallows (1987) documented the rise of watertables that was gradual for the first decade and then accelerated to a rate of 1.6 m/year. It was predicted that if current irrigation continued, equilibrium would be attained by the end of the C20th century. Shallow groundwater levels within the CIA have stablised since the early 1990s, with a much slower rate of rise and new dynamic equilibrium apparent at between 2–4 m depth (CIC 2000).

In 1999, the Murrumbidgee Groundwater Management Committee (MGMC) compiled a discussion paper of groundwater issues. In accordance with the NSW Groundwater Quality Protection Policy (DLWC 1998b), it recognised that groundwater quality protection should be integrated with the management of groundwater quantity. No significant groundwater contamination was known to have occurred in the Lower Murrumbidgee - this, it was recognised, was due partly to the fact that relevant data had not been examined. The lack of variation in groundwater quality with depth was thought to indicate a lack of movement between aquifers. Leakage through abandoned & improperly constructed bores was flagged as an issue requiring attention.

Little regional scale hydrochemical information is available for this area, with the exception of a detailed hydrochemical study of shallow groundwater (Williams & Ward 1987), and a study of aquifer residence times by Drury et al. (1984). Although the Lower Murrumbidgee aquifers have been nominated as the most highly vulnerable inland aquifer in New South Wales (DLWC 1998a), the extent of nutrient and agrochemical contamination has not been assessed.

1.6 The Tubbo site

Several criteria were used to select the primary study site including: an area in the early stages of groundwater development where the saturated zone was still at >20 m depth; an existing high volume irrigation bore; ease of access during wet weather; and a cooperative and supportive landholder.

The Tubbo Irrigation property (the eastern part of historic Tubbo Station) was located near the origin of western and southwestern flow paths (Figure 1.2B,C) at the apex of the Murrumbidgee alluvial fan. Since this area lies outside the surface channel supply of the Coleambally irrigation area, groundwater is used to irrigate pasture for livestock although neighbouring properties irrigate rice and, on sand dune country to the east, vegetables (Figure 1.3).

At this site, 6 nested piezometers were installed at depths between 23 and 95 m depth, located 35 m upgradient from a large volume irrigation bore 59113 (Figure 1.4, Table 1.1). Construction details including placement of bentonite seals is included in Appendix B. Unfortunately, due to problems encountered in the field, a bentonite seal was not placed below the B3 piezometer screen in the Lower Shepparton silt unit (M. Groskops, pers. com.). Since this piezometer was installed in the same drill hole as B6, upwards annulus flow was possible from the underlying aquifer, such that data from this piezometer were not reliable.

The physical characteristics of the aquitard were investigated by borehole and surface geophysical techniques and analysis of minimally disturbed cores. Groundwater levels were monitored intensively during the first pumping cycle and then every 6 hours between August 1998 and March 2001 in the nested piezometers and 2 DLWC observation bores, cited at distance of 1.8 km from the pumping bore. Key hydrochemical and isotopic indicators were sampled before, at several stages during, and after the 1998–99 irrigation season.

Photographs of the Tubbo irrigation and monitoring bores are provided at the end of Chapter 1 (Figure 1.6).

1.7 Sketch of this thesis

Chapter 1 of this thesis included general background; unknowns; the environmental setting and overview of study sites; hypotheses and conceptual model; and project scope and objectives.

In Chapter 2, the environmental setting of the study area in the Lower Mur-

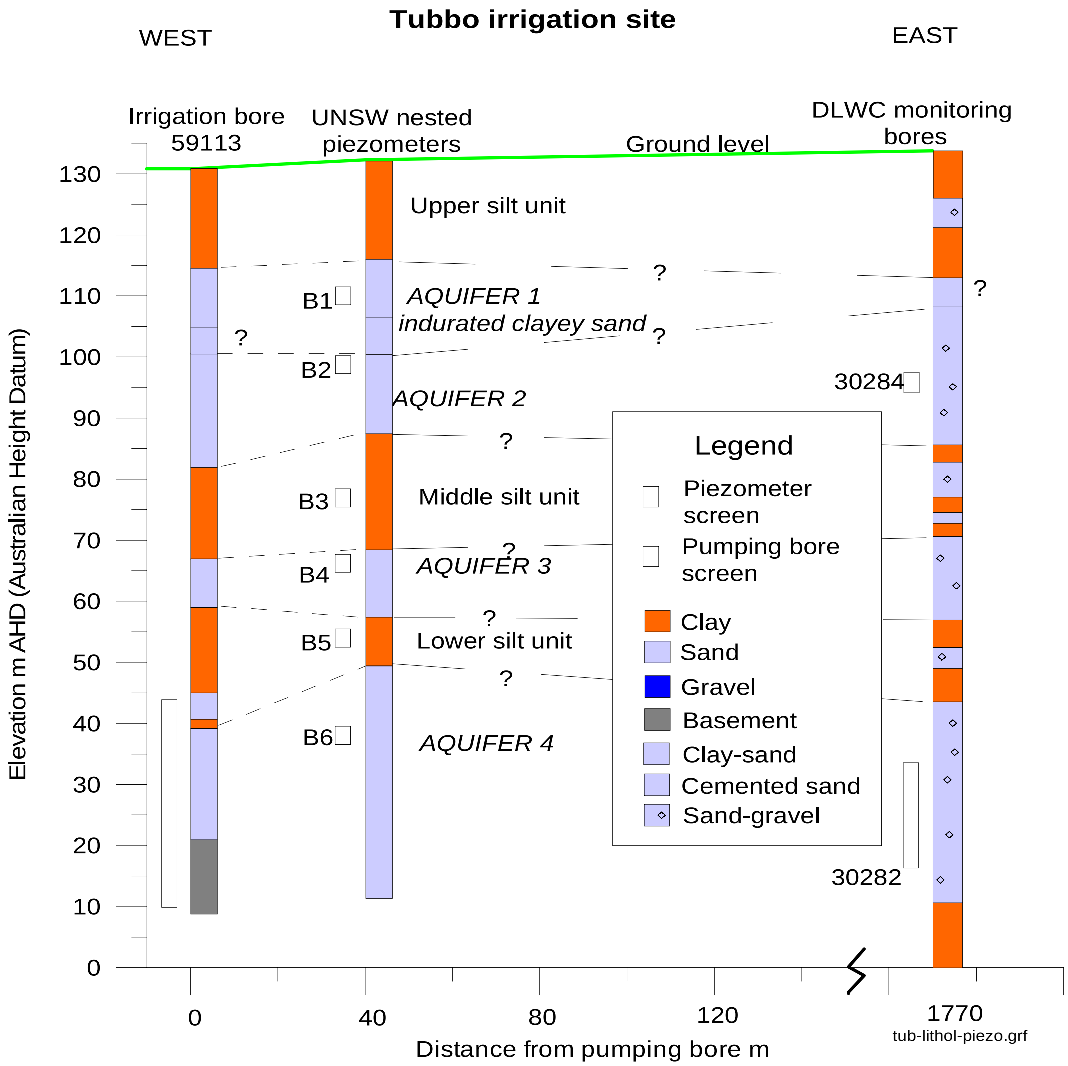


Figure 1.4: Aquitard and aquifer geometry and piezometer installations at the Tubbo site (see Figure 1.3 for piezometer locations).

Table 1.1: Piezometer installation details at Tubbo site.

Bore	Screen depth (m)	Mid screen depth (m)	Drilled depth (m)	Elevation top casing (m AHD) [a]	Casing height (m)	Casing ID (mm)	Install date	SWL Feb-99 (m bc) [b]	Purging 1 bore vol (L)
UNSW nested piezometer site, E427435 W6161840									
B1	21-24	22.5	80	130.964	0.432	50	Jun-98	17.46	15.4
B2	32-35	33.5	69	130.933	0.4	50	Jun-98	18.08	–
B3	54-57	55.5	96	130.964	0.43	50	Jun-98	18.32	77
B4	65-68	66.5	69	130.933	0.4	50	Jun-98	18.28	101
B5	77-80	78.5	80	130.964	0.432	50	Jun-98	18.29	123
B6	93-96	94.5	96	130.964	0.43	50	Jun-98	18.32	155
Tubbo irrigation bore, E427410 W6161865									
59113	86.8-120.7 [c]	103.75	128	131.0	–	406	Jan-84	18.3	13260
DLWC observation bores, E429180 W6161900									
30284	36.2 - 39.5	37.85	39.5	133.66	0.52	203	May-72	18.1	682
30282	100.2 - 117.2 [d]	108.75	138.6	133.75	0.66	117	May-72	18.04	1130

[a] m AHD = metres Australian Height Datum
[b] bc = below steel casing
[c] 5 screen intervals
[d] 2 screen intervals

rumbidgee is described, and the state of local knowledge pertaining to aquitards, diffuse recharge and groundwater quality is reviewed.

Chapter 3 comprises an international literature review of leakage mechanisms through clayey aquitards and the significance of groundwater quality variability.

Chapter 4 describes field and laboratory investigation techniques.

Chapter 5 documents geophysical and permeability characteristics of each aquitard including the spatial distribution of the near surface deposit.

Chapter 6 examines the changes in groundwater hydrology induced by groundwater pumping and quantifies vertical leakage by developing a numerical axisymmetric model of groundwater flow near the Tubbo irrigation bore.

Chapter 7 highlights the hydrochemical and isotopic variability, both at a regional scale over several decades, and at a local site scale during the irrigation season. Mass balance hydrogeochemical modelling accounts for mixing and other processes.

Chapter 8 synthesises hydrogeology, hydrodynamics and hydrogeochemicstry in an integrated assessment of the relative importance of solute transport mechanisms.

Finally, Chapter 9 presents a series of conceptual models which were developed for aquifer-aquitard systems, and identifies key research findings and future challenges.

Publications and reports arising from this project, together with supplementary datasets and interpretations are archived on the accompanying CD-ROM.

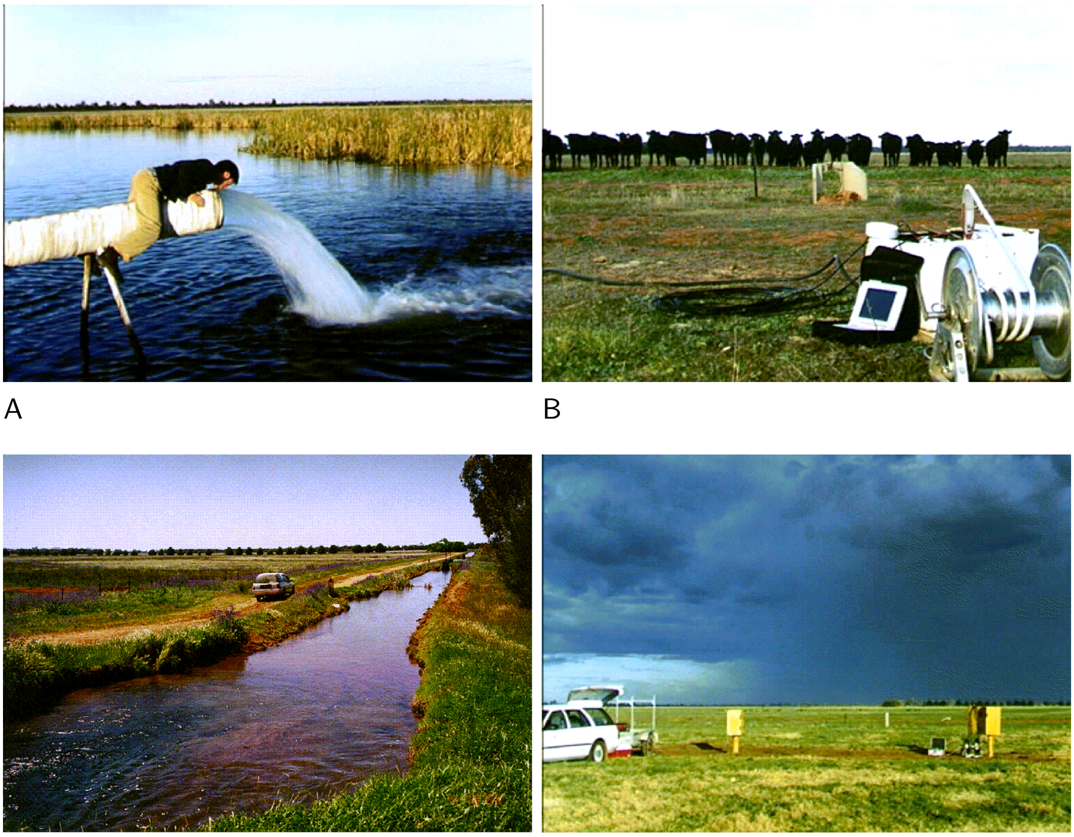

A


B


C


D

Figure 1.5: Surface and groundwater resources supplying agriculture in the Lower Murrumbidgee area. **A** Flood irrigated rice crop. **B** Bore drain distributing groundwater from the pump head. **C** Groundwater bore discharge into main Coleambally irrigation supply channel at Tubbo south. **D** Outlet of main Coleambally pump bore that was installed in 1990 for mitigation of shallow watertables.



A B C D

Figure 1.6: The Tubbo site. **A** Discharge of groundwater bore into temporary storage dam. A single irrigation bore such as this one, pumping at 280 L/s, would be capable, at that rate, of supplying drinking water to all Australians. **B** Nested piezometers installed near the irrigation bore in aquifers and aquitards between 23 and 95 m depth. **C** Channel supplying groundwater for flood irrigation of pasture. **D** DLWC monitoring bores, located 1.8 km upgradient from the irrigation bore.

Chapter 2

MURRUMBIDGEE GROUNDWATER ENVIRONMENT

This chapter provides a regional context for the study by describing climate, hydrology and geological history of the Murrumbidgee area. The nature of the groundwater resource are considered, along with the impacts of development with particular emphasis on predicted and observed water quality changes.

2.1 Climate

The modern climate will be briefly described, although it is climatic change over geological time which has the greater bearing on salinity.

This area of inland Australia is characterised by a semi-arid climate, with hot summers and low, generally uniform rainfall. Median annual rainfall at Leeton is 434 mm, with the annual rainfall range of 157 to 822 mm since records began in 1914 (Bureau of Meteorology data). Figure 2.1 shows average annual rainfall between 1914 and 2000. Slightly higher rainfall occurs during winter which also has more wet days.

Residual mass curves of rainfall, relating rainfall trends to long term monthly averages, indicate increasing rainfall since the mid 1980s (Lawson 1992). Rainfall decreases slightly from east to west across the study area, and some variability was observed between rainfall recorded at Leeton, Griffith and Wagga Wagga during the study period.

Annual average Class A pan evaporation is 1 815 mm per year at Griffith (Lawson 1992). Evaporation is greatest in summer with 270 mm per month, with a low of 50 mm per month in winter. Potential evaporation thus exceeds rainfall by a factor of about 4.

During the study period, rainfall was 366 and 553 mm during 1998 and 1999 respectively. Monthly rainfall for Sep-98, Jan-99, Mar-99 and May-99 was above the median

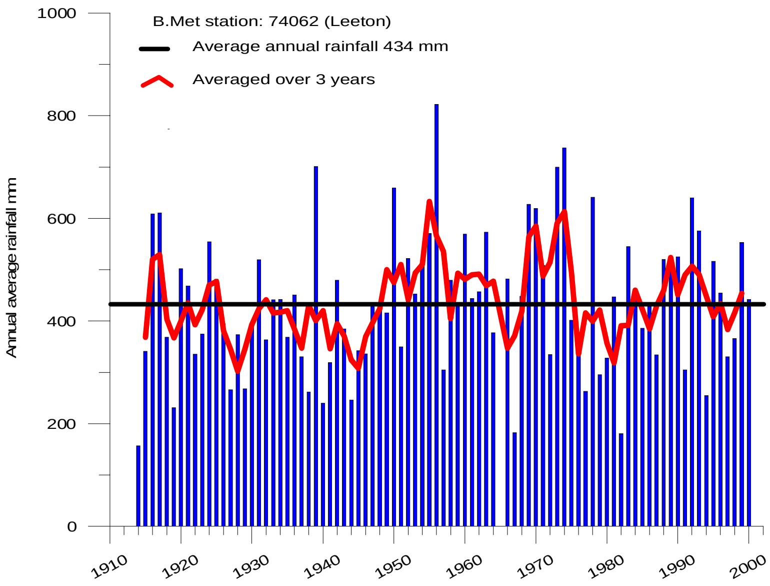


Figure 2.1: Average annual rainfall at Leeton, 1914–2000.

monthly rainfalls.

2.2 Hydrology

2.2.1 Modern surface drainage

The Murrumbidgee River drains the northern part of the Snowy Mountains and enters the eastern edge of the Murray Basin at Narrandera, about 30 km east of the primary study site Tubbo (Figure 1.2). The modern Murrumbidgee River, established about 12 000 years ago, is highly sinuous, and has a very low channel gradient of 1:7 500. The Lower Murrumbidgee alluvial fan is also of very low gradient, situated at an elevation 300 m above sea level, and >3 000 km by channel to the Indian Ocean.

A relatively low proportion of precipitation is discharged by the rivers of the Riverine Plain (estimated to be 4% in the Murray Basin by Simpson & Herzceg 1991). Further details about runoff variability within the catchment, flow regulation and water use in the Murrumbidgee Valley is provided by DWR (1993).

The Murrumbidgee River is thought to be in hydraulic continuity with the lower sand and gravel aquifer, particularly in the vicinity of the alluvial fan apex. The lower sand and gravel zones therefore form an important regional aquifer system that is recharged by leakage from the River. It is significant that the hydraulic head in the river is maintained at an elevated level during the summer irrigation season by release of water stored

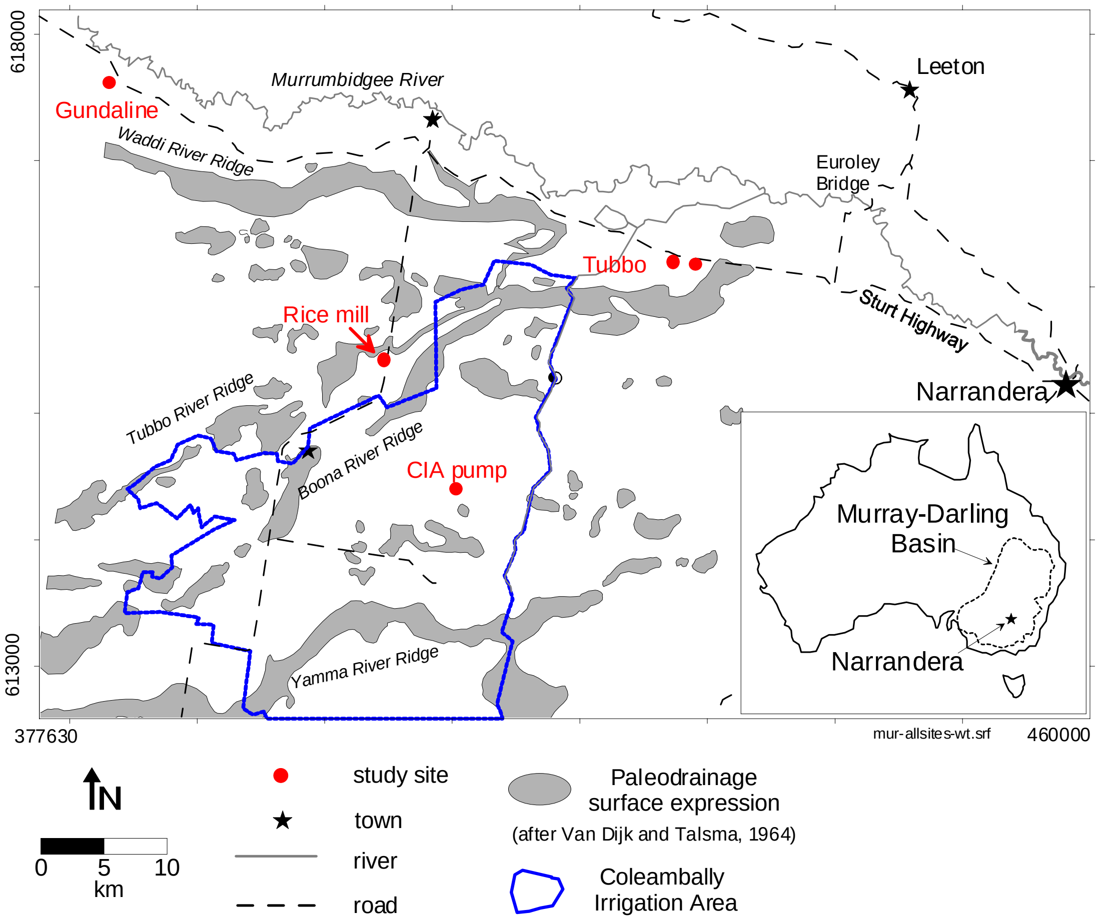


Figure 2.2: Palaeodrainage and interchannel plains relative to study sites.

in the Burrunjuck Reservoir, which is located in the foothills of the mid-Murrumbidgee catchment.

2.2.2 Palaeodrainage

The form of the Murrumbidgee River has varied greatly during Pliocene-Pleistocene time, resulting in a highly complex depositional jigsaw of fine and coarse grained deposits. The palaeodrainage distribution of van Dijk & Talsma (1964) was based on the occurrence of channel sands as outcrop (Figure 2.2). The outcrop represents the 'tip of the iceberg' so to speak, since a more complex 3 dimensional palaeodrainage system extends to bedrock. palaeodrainage channels would appear in plan view as a beaded belt, embedded in a three dimensional, though non-laterally extensive matrix of clay (Galloway & Hobday 1996). Coarse channel fill sediments, in this form of river deposit, would account for between 5 and 20% of the total volume.

Fluvial deposits on the Riverine Plain reflect evolving climatic and hydrological regimes. McNally & Sutherland (1997) synthesised the findings of early workers who relied on soil survey and aerial photography (Butler 1950; van Dijk & Talsma 1964; Pels 1964; Schumm 1968) with the refinements of Fried (1993), Page & Nanson (1996), Page et al. (1996) and Page et al. (1991). Three distinct styles related to palaeoclimatic conditions have been recognised as follows:

- Prior streams - 100 to 50 ka BP, wide braided, straight, bedload dominated
- Ancestral rivers - 50 to 10 ka BP, very wide meandering mixed load, marked by dune levees and meander scrolls. Discharge at Darlington Point estimated to be 5 times modern volumes.
- Modern streams - since ∼12 000 BP, extremely sinuous with no distinct levees or dunes.

The modern landscape was formed during the 'Prior Stream Era' associated with falling sea levels (Hallows 1987). An internal drainage system existed for hundreds of thousands of years whilst dunes blocked access to the sea. Streams emerged from the mountains to deposit sediments on the plain. The prior stream channels being coarse grained were leached of salt, but the fine grained interchannel sediments were not. The final 40 000 years (Upper Pleistocene epoch) has seen the rivers once again reconnect with the sea and superimpose their alluvial beds on the prior streams.

Poor understanding of fluvial geomorphology led to the construction of irrigation supply channels on the relatively elevated prior stream beds. Consequently, many channels are prone to considerable leakage.

2.3 Geological setting

2.3.1 Basement geology

Basement rocks in the Lower Murrumbidgee area consist of Paleozoic and Mesozoic rocks (van der Lelij et al. 1987). Near Narrandera, bedrock geology is dominated by Ordovician metasediments of quartzite, slate and schist. Small Silurian granite intrusions occur beneath the CIA (Lawson 1992). These metasediments are overlain by residuals of flat lying Devonian sandstone in the Narrandera area (Jankowski et al. 1994).

2.3.2 A brief history of the Murray Basin

The Riverine Province of the Murray geological basin, contains Cainozoic sediments of mainly continental origin (Figure 2.3). Marginal marine units interfinger the Renmark Formation on the western Mallee boundary, but do not occur near the study area. Sediments have accumulated on the Lower Murrumbidgee alluvial fan since the mid-Tertiary period.

The edge of the Basin has acted as a hinge zone without significant subsidence or sediment accumulation, such as occurred further west. As a result, there are only approximately 120 m of unconsolidated clays, sands, silts and gravels overlying the pre-Cenozoic basement rocks beneath the alluvial fan.

A shallow sea that invaded the Murray Basin approximately 30 Ma years ago, deposited a marine sequence of clays and limestone . To the east of this sea a zone of deltas and swamps was formed within which silts, sands, clays and coal accumulated. These deposits are known locally as the Renmark Formation and occur beneath the more western parts of the fan.

Sand and gravel lenses occur most frequently in the lower part of the succession and are thought to have accumulated between 6 and 2 Ma years ago as a sand sheet deposited by the proto-Murrumbidgee (Figure 2.4). In the central-west of the Murray Basin, the Pliocene sand deposits are overlain by a thin skin of clay (Blanchetown Clay) deposited in a large former lake system (Lake Bungunnia). The lake dried up approximately 0.5 Ma years ago and this marked the change to drier conditions associated with the later part of the Pleistocene. The Pliocene sands are locally called the Calivil Formation in the eastern part of the Basin.

Pleistocene and Holocene sediments that have accumulated over the Calivil Formation are more clay rich although major sand lenses also occur. These deposits are locally referred to as the Shepparton Formation. There is no clear lithological difference between the upper parts of the Renmark, the Calivil or the Shepparton Formations, other than the appearance of a generally fining upwards sequence. The first occurrence of coal is often assumed to mark the top of the Renmark Formation.

2.4 Aquifer systems

Major aquifers of the Murray Basin are located within the Renmark Group, Murray Group, Pliocene Sands and Shepparton Formation (Evans & Kellett 1989). The Murray Basin is a closed groundwater basin with no outflow apart from discharge to playa lakes in the

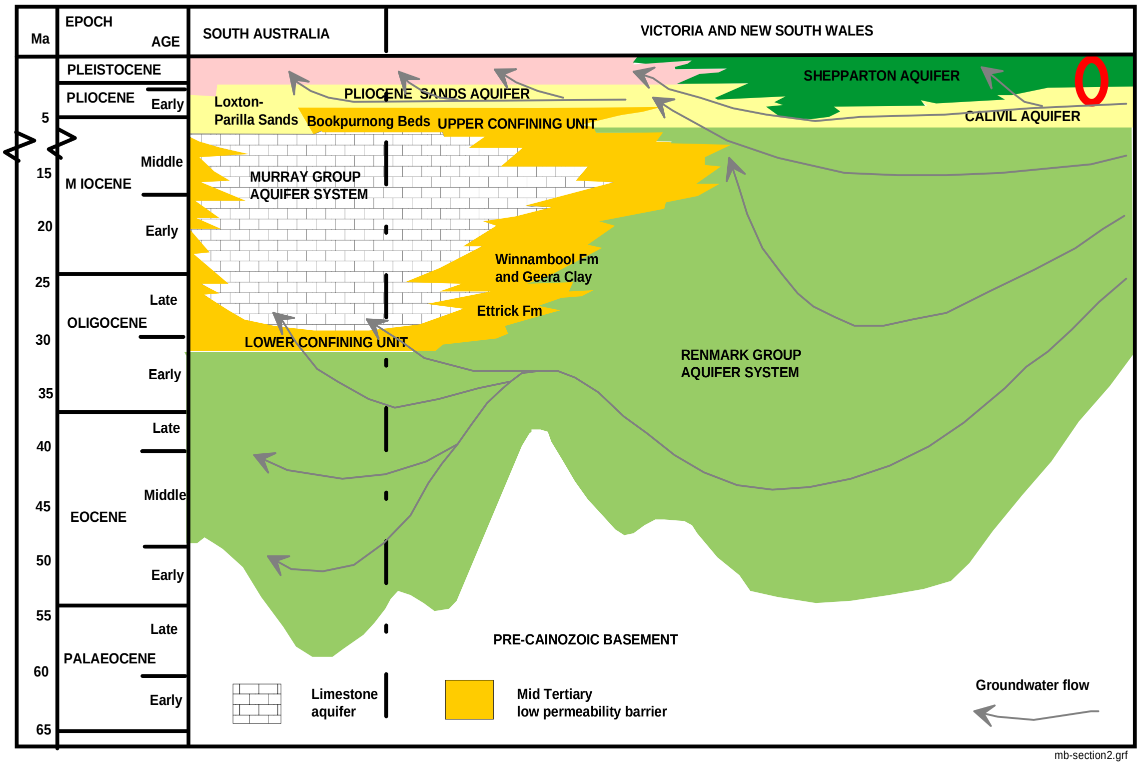


Figure 2.3: Murray Basin groundwater section (Brown, 1989). Project focus marked by **O**.

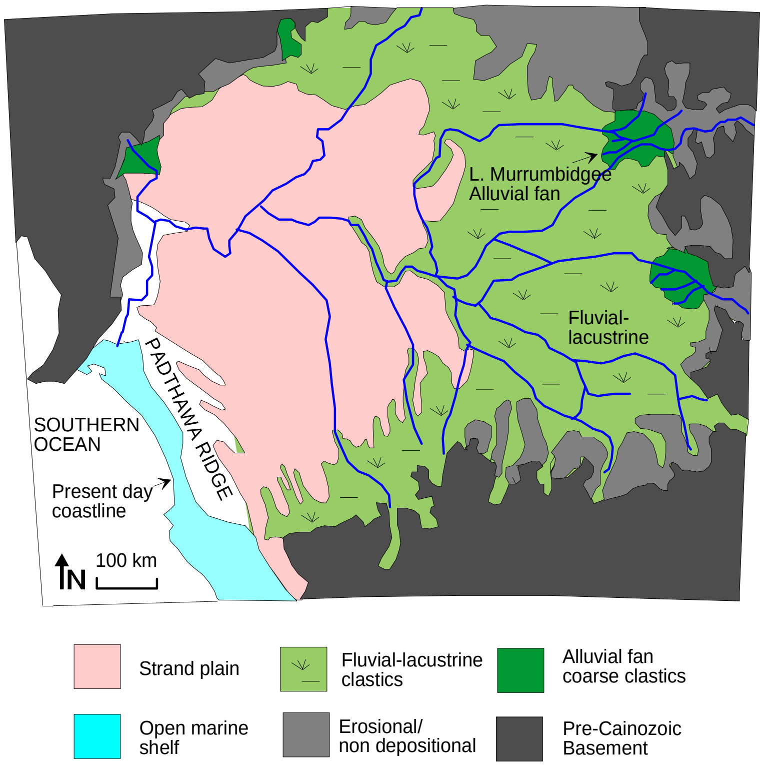


Figure 2.4: Palaeofacies of the Murray Basin depicting development of alluvial fans at the margins of the Riverine Plain - Pliocene period of Calivil Formation and Lower Shepparton Formation deposition (after Brown, 1989).

western Mallee area.

2.4.1 Renmark Formation

The Renmark Formation is a continuous basal confined aquifer group of Tertiary to Middle Miocene (50 to 15 Ma BP) age (Lawrence 1975). It is characterised by terrestrial deposits of fluvial clay, silt, sand, minor gravel and ubiquitous carbonaceous deposits. The Lower Renmark Group infills tectonic infrabasins in the Riverine and other provinces of the Murray Basin (Evans & Kellett 1989). The Upper Renmark Formation is distinguished from the Middle Renmark Formation by finer grained and carbonaceous-lignitic deposits, although at the fringes of the basin, such subdivisions are tenuous as these deposits are homogeneous. These two formations were referred to as the Olney Formation and Warina Sand respectively (Lawrence 1975), terminology that is no longer in common usage.

Horizontal hydraulic conductivity in the Renmark aquifers averages 3 m/day on a regional scale, but is up to 100 m/day within the alluvial fans (Evans & Kellett 1989). Rythmic layering of coarse and fine sediments suggest that vertical hydraulic conductivity is at least an order of magnitude less than horizontal hydraulic conductivity. Lawrence (1975) suggested this ratio to be 20:1.

Groundwater quality within the Renmark aquifer system is variable. The dominant watertype is NaCl throughout, with distinctive increases in salinity in the west, below the Loxton-Parilla sands and Geera clay. Groundwater is generally slightly acid to neutral but may be highly reduced in the vicinity of lignitic sediments.

2.4.2 Calivil Formation

The Calivil Formation is part of the Pliocene Sands aquifer system of the Murray Basin, which also includes the Loxton-Parilla sands in the west. Sands and gravels of the Calivil Formation were deposited by the ancient Murray River drainage system during the Late Miocene and Pliocene eras (15–5 Ma BP). The coarse clastic deposits, particularly in 'deep leads' of highland valleys, were deposited as backfill due to a rise in base sea levels during a late Pliocene transgression. In the New South Wales portion of the Murray Basin, the Calivil Formation is 35–70 m thick, and directly overlies bedrock at the basin margins (Woolley & Williams 1978).

Woolley & Williams (1978) described the Calivil Formation in the apex of the Lower Murrumbidgee fan as being dominated by pale grey coarse sand and fine gravel, containing pale grey to white kaolinitic clay and very thin bands of carbonaceous clay. Lawson (1992) estimated that sand and gravel layers comprised 50–70 % of the Calivil Formation. The top of the unit is generally defined as the uppermost occurrence of grey quartzose sand or white kaolinitic clay (Lawson 1992).

The Calivil Formation directly overlies and is in hydraulic continuity with the underlying Renmark aquifer systems, and is semiconfined by the Shepparton Formation above. Recharge is controlled largely by the permeability of the confining Shepparton Formation and is greatest in the eastern fringes of the basin. Transmissivity of up to 6 500 m^3/day/m near Darlington Point was reported by Wolley and Williams (1978). It is the most productive aquifer within the Lower Murrumbidge area with the largest yielding bores in Australia located near Darlington Point.

2.4.3 Shepparton Formation

The Shepparton Formation, which is of Late Pliocene to Pliestocene age, directly overlies the Calivil Formation and interfingers with the Loxton-Parilla sands to the west. It is a complex assemblage of variegated clay and lenses of yellow and brown polymitic sands which were deposited in a fluvio-lacustrine environment.The shoestring shape sand lenses are mainly restricted to the uppermost 30 m of the sequence which may be up to 95 m thick (Woolley & Williams 1978). In the Coleambally area, sand lenses comprise about 20–30% of the profile, particularly in the top 25 m and between 45–70 m depth (Lawson 1992).

Due to the predominance of clay, regional hydrualic conductivity is about 2–3 m/day and the formation is of limited water yielding capacity. However, Evans & Kellet (1989), estimated hydraulic conductivities of sandier zones between 25–100 m/day and averaging 45–50 m/day. Lawson (1992) cites DWR estimates near Darlington Point of 50–60 m/day.

2.5 Aquitard deposits

Distinguishing characteristics of aquitard deposits described in the south-eastern Murray-Darling Basin are given in Table 2.1. The Pliocene-Pleistocene aquitards are distinctive from earlier marine deposits, comprised of clay interbed deposits. Previous knowledge of the Calivil aquitard was restricted to its clay mineralogy, which was identified as grey to white kaolinitic clay (Woolley & Williams 1978). Far more is known about clay within the Shepparton Formation, particularly near the surface where road cutting exposures and soil profiles have been investigated by a number of workers.

2.5.1 Mid-Tertiary permeability barriers

Regional groundwater flow in the Murray Basin is influenced by mid-Tertiary permeability barriers which include the Ettrick Formation, the Winnambool Formation, the Bookpurnong beds and the Geera Clay (Brown & Radke 1989) (Table 2.1). The Geera clay is the most important regional aquitard which has a profound effect on pressure distribution and water chemistry because it forms a 100 km wide permeability barrier through the middle of the Murray Basin (Figure 2.3). It is a rare example of detailed field investigation of an aquitard in the Murray-Darling Basin (Table 2.1). The Geera clay was deposited by a marine transgression during the Oligocene-Miocene period (12–32 Ma BP), in an arcuate zone in the west central Murray Basin (Brown & Radke 1989).

The low permeability Geera clay causes flow lines of the Renmark Formation to diverge and precludes overturn of highly saline groundwater at shallow depths (Hanor 1987). Step wise salinity increase in the Renmark aquifers below the Gerra clay reflect downward diffusion of connate salts (indicated by downwards decrease in concentration gradient). Jones et al. (1994) found that the Upper Geera clay was acting as a sink for saline groundwaters diffusing downwards from the overlying Loxton-Parilla sand. Connate waters in the Lower Geera clay however, were a source of salt diffusing downwards. Hanor (1987) considered that the continuous variation in chemical composition indicated continuous vertical hydraulic continuity. Downwards diffusive flux, which dominated advective transport, was estimated to be 1 g NaCl/m^2/year.

2.5.2 Soils

The distinction between soil and alluvial deposit is difficult in fluvial environments where soil horizons are often not well developed (Charman & Murphy 1991). Consequently, soil investigations, which describe similar subsurface materials (apart from organic matter) are included in this discussion. The first assessment of soil properties in the Lower Murrumbidgee area was a 1954 reconnaisance for the proposed CIA development, reported by van Dijk & Talsma (1964). Numerous subsequent studies investigating salinity characteristics and hydraulic properties were recently summarised by Hornbuckle and Christen (1999).

Soil types in the Lower Murrumbidgee floodplain, which account for 75% of the landscape (van der Lelij 1989), include transitional red-brown earths and self-mulching clay, interspersed with grey and brown soils of heavy texture. Prior drainage levees are characterised by sandy soils, red-brown earths and solodized solonetz soils. Differentiation

Table 2.1: Aquitards in the south-eastern Murray Darling Basin.

Unit	Age	Basin area	Comments
Shepparton & Calivil interbeds [a]	Pliocene-Pliestocene	E	Thin, discontinuous, overbank deposits
Blanchetown Clay [b]	Pliestocene	W	Thin, discontinuous lacustrine deposits
Bookpurnong Beds [c]	L. Miocene (∼6 Ma)	W	Shallow marine calcareous clayey silt. Directly overlies Geera clay and Winnambool Fm in some areas, ∼50 m thick.
Geera Clay [d]	L. Oligiocene to M. Miocene	N&E	Shallow marine silty mud forms most effective permeability barrier. ∼45 m
Winnambool Fm [e]	Miocene (15–25 Ma)	Central	Shallow marginal marine deposit causing regional discharge. Includes minor channel sands now cemented. ∼37 m thick
Ettrick Fm [f]	Oligocene (25–30 Ma)	SW	Shallow marginal marine calcareous clay, permeable, allows fresh waters to recharge Murray Gp aquifer. ∼20-30 m thick

[a] Timms & Acworth (2002b), Timms et al. 2000, Woolley & Williams 1978
[b] Stephenson 1986
[c] Brown 1989
[d] Hanor 1987, Brown & Radke 1989, Jones et al. 1994
[e] Brown & Radke 1989
[f] Brown & Radke 1989

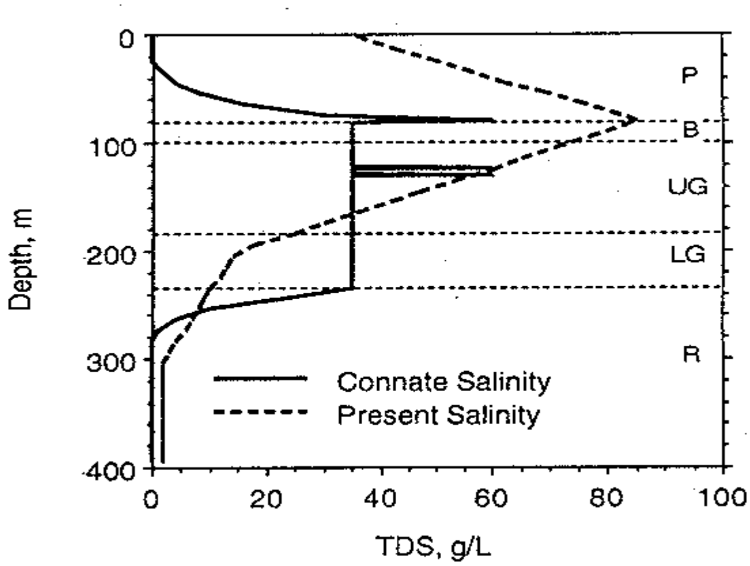


Figure 2.5: Comparison of estimated connate and present pore-water salinities of aquifer-aquitard sequence at Piangil West site. (After Jones et al. 1994) P is Parilla Formation, UG and LG are Upper Geera and Lower Geera respectively and R is Renmark Formation.

of soil types is largely the result of drainage differences (Butler & Hutton 1956). Many soils are highly weathered, marked by the absence of mica in the top metre of the profile.

Butler (1950) studied the link between drainage and soil distributions in regard to salinity. Introducing the concept of original salinity, it was postulated that the spatial distribution of both fine sediments and salinity increased with distance down a prior stream channel and with distance transversely from the prior stream channel. In this conceptual model, sources of salt, from mountain weathering and aoelian deposition were redistributed by prior streams and the original salinity pattern subsequently overprinted by flooding, runoff and leaching.

2.5.3 Aeolian contributions

Butler & Hutton (1956) introduced the term parna as an aoelian clay related to arid climatic phases in the Quaternary. The Lower Murrumbidgee area is within the east 'Australian dust mantle', which deposited aeolian clay known as the Widgelli parna in an area west of Darlington Point and eastwards to Wagga. High dunes between the Tubbo site and Narrandera, and close to Wagga Wagga are also attributed to aeolian sedimentation (eg. dunes in the lower right of the aerial photograph shown in Figure 1.3 where center pivots are seen to irrigate vegetable crops). The higher proportion of clay in the southern and western areas of the Lower Murrumbidgee has been attributed to aeolian deposits.

Aeolian deposits are characterised by higher clay content, sub-plastic behaviour and the absence of lime which has been leached deeper in the profile (Butler & Hutton 1956). Parna was deposited as aggregates of clay together with sand particles of a similar size known as companion sand. Originally, parna deposits contained between 3–20% carbonate, with a lesser quantity at the eastern margins of the Riverine plain, though much of this carbonate is considered to have been leached, contributing to the formation of secondary carbonate (Butler & Hutton 1956).

Subplastic behaviour, resulting in an underestimation of clay content, has been associated with aeolian deposited clay (Beattie 1970). Clays which resist dispersion by classical methods are well known, though the nature of internal cementation which apparently causes the phenomena is not known (Butler 1976; Beattie 1970).

2.5.4 Soil permeability

Direct measurement of subsurface hydraulic conductivity, K, on a site and field scale is limited to soil infiltration studies of the uppermost 3 m of the profile (Talsma & van der Lelij 1976; Talsma & Flint 1958). The most effective barrier to downward percolation is known to be the heavy clay B horizon of duplex transitional red-brown earths (van der Lelij 1989). Permeability determined on soils is comparable to K_v values for the Upper Shepparton aquitard (Table 2.2).

Talsma & Flint (1958) studied the variation of hydraulic conductivity of soils in the Murrumbidgee irrigation area with degree of cementation and depth. Hydraulic conductivity was found to decrease with depth to 2.5 m, particularly in heavy-textured clay, and to decrease with increasing clay and silt content. The size and shape of sand within a clayey silt was not expected to affect hydraulic conductivity except where clay content was less than 15%. An impermeable layer was defined by hydraulic permeability $\leq$0.1 of the overlying layer.

Hydraulic conductivity increased with increasing cementation due to restructuring associated with intra-particle formation. Subplastic soils that were weakly, moderately and highly cemented were characterised by K of 5.2×10^{-6}, 1.4×10^{-5}, 3.0×10^{-5} m/s respectively. By comparison, the K of hardpan material where the total pore space was filled by cementing material was about 8.6×10^{-6} m/s.

Talsma & van der Lelij (1976) studied infiltation and water movement in a vertisol soil below a rice paddy on the Lower Murrumbidgee. The soil profile was 62% clay, mostly in the form of swelling montmorillonite. Standard ring infiltrometers were used to mea-

Table 2.2: Vertical hydraulic conductivity of soil and aquitards in the Murray Basin and Lower Murrumbidgee areas.

Unit	K_v (m/s)	Comments
Shepparton	2.3×10^{-9}–3.5×10^{-8}	7 day pump test in Calivil Fm near Shepparton, Vic. [a]
	4.4×10^{-8}–5.0×10^{-8}	30 day pump test in Calivil Fm near Elmore, Vic. [b]
	6.9×10^{-10}–4.6×10^{-7}	lower part of Shepparton during 468 day CIA pump test [c]
	1.2×10^{-9}	average
	1.2×10^{-5}	estimate used for saline groundwater mixing (1 m/day) [d]
	$8.6\times10^{-}$	hardpan cemented subsoil (Talsma & Flint 1958)
	2.4×10^{-8}	mean of ring infiltrometer tests (Talsma & van der Lely 1976)
	2.4×10^{-8}	water balance tests (Talsma & van der Lely 1976)
Low permeability matrix	1×10^{-8}	definition [e]
Clay liner	1.2×10^{-9}	Canadian regulations for constructing wastefill clay liners [f]

[a] Olshina 1988
[b] Brinkley & Reid 1990
[c] Lawson 1992
[d] Williams 1997
[e] Neuzil 1986
[f] Johnson et al. 1989

sure soil K of between 6×10^{-9} and 1.2×10^{-7} m/s, with a geometric mean of 2.4×10^{-8} m/s (2 mm/day). These values contrasted with the wide range of values derived from piezometer measurements from practically impermeable to 4.4×10^{-7} m/s. Independent estimates using the water balance method for the same field yielded a K of about 2.4×10^{-8} m/s (Talsma & van der Lelij 1976).

2.6 Groundwater resources

The groundwater resource within the Lower Murrumbidgee alluvial fan is of immense value for agriculture, local town water supply, river base flows, and as a strategic national water reserve. The latter point is particularly pertinent for provision of potable water for future generations on the driest inhabited continent on earth. Balancing usage and recharge is

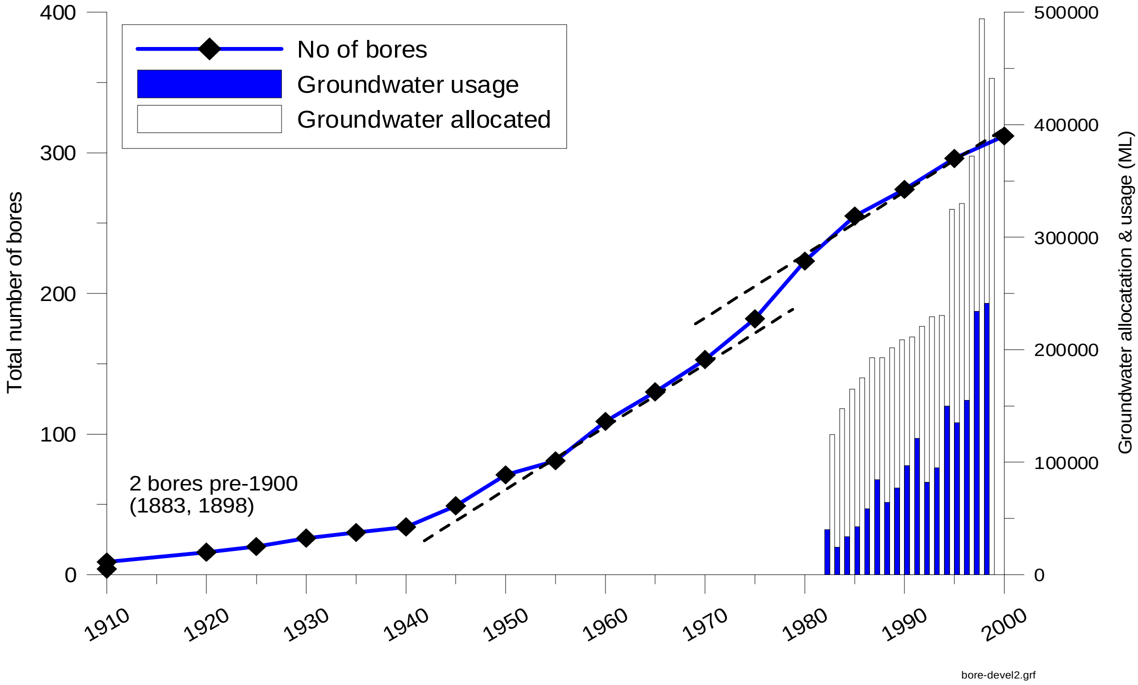


Figure 2.6: Groundwater development in the Lower Murrumbidgee study area showing near steady rate of increase in the number of bores between 1940–2000, and increased allocations and usage since the mid 1990s (DLWC data, P. Kumar & L. Webb, pers. com.).

critical to sustaining this resouce.

Local groundwater resources in terms of yield, salinity and bore development are summarised by the Narrandera 1:250 000 hydrogeological map (Woolley 1991). The highest yeilding and lowest salinity bores are as expected, located adjacent to the Murrumbidgee River and Yanco creek.

2.6.1 Groundwater levels related to usage and allocation

Groundwater development began in the 1850s with timber shored hand dug wells up to 20 m depth (Kelly 1978). The need for stockwater on dry properties away from the River meant that by the 1960s one was 'never out of sight of a windmill' (Ronald 1960). Windmills were generally 20–50 m deep. By 1908, steam driven drilling machines were capable of screening off shallow saline groundwater to reach deeper fresh waters. Bore technology capable of pumping from deep multi-layered aquifers became available in the 1970s, some time after surface water irrigation supplies were established. The rate of bore drilling was greatest in the mid-1980s, although the rate of bore construction was otherwise constant between 1940–2000 (Figure 2.6).

Lawson & Webb (1998) provide detailed information on allocation and groundwater usage distribution, the highlights of which are included here. Most groundwater usage in the Lower Murrumbidgee valley today is for irrigation, with 2% for town water supply and industrial use. Groundwater allocations amounted to 494 000 ML in 97/98, although less than half (241 000 ML) was actually used. Despite an embargo on new allocations in 1987, usage of existing allocations is steadily increasing every year.

In the Lower Murrumbidgee, 10% of estimated recharge is reserved for environmental flows, while the balance of 225 000 ML is considered to be the sustainable yield. Current usage therefore exceeds sustainabile yield by ~6 000 ML, and if all allocated groundwater was to be pumped, the sustainable yield would be greatly exceeded.

Lawson & Webb (1998) note that groundwater level declines are occuring in some areas of the Lower Murrrumbidgee alluvial fan. Decreasing trends have emerged relatively recently, and are restricted to areas outside the CIA, particularly west of Darlington Point. Groundwater level declines in the deep aquifer became evident in the mid-1990s. In 1997–98 seasonal drawdowns of up to 16 m were documented, and long term drawdowns of up to 6 m occurred. By contrast, equilibrium or slightly rising long term groundwater levels were noted for the deep aquifer in some areas of the CIA and areas distant from the river and groundwater pumping (Lawson & Webb 1998) .

2.6.2 Recharge sources

The total volume of low salinity water in storage in the Lower Murrumbidgee valley is estimated to be 2 000 Ma ML (Lawson & Webb 1998). The large groundwater storage, and the limited vertical connection to the confined aquifer mean that individual recharge events such as major flooding do not result in major groundwater level rise.

Sources of recharge are quantified in Table 2.3. River recharge is estimated at 180 000 ML/year or 80% of total recharge. Diffuse recharge of rainfall and irrigation accounts for the remaining 30 000–70 000 ML /year (Lawson 1996), although water balance studies suggest diffuse recharge as high as 96 800 ML/year (van der Lelij 1989). The 96 800 ML/yr deep drainage was calculated as the difference between nine year average rice water consumption (for 1975 to 1984) and the annual crop requirement of 15 ML/ha (van der Lelij 1989). This means that about 20% of surface channel water which is supplied to the CIA is lost to groundwater by seepage.

River recharge

Drury et al. (1984) found that most recharge on the Riverine Plain occurred from the Murrumbidgee River. River losses have been estimated at 1 ML /day/km of river between Narrandera and Darlington Point (Woolley & Kalf 1981) where hydraulic continuity with the lower sand and gravel aquifer of the Calivil Formation is greatest. Lateral groundwater flow rates in the deeper aquifer are 7–10 m/year away from the river, though in the absence of pumping the gradient driving flow is very small. It is significant that the hydraulic head in the river is maintained at an elevated level during the summer growing season by release of water stored in the Burrunjuck Reservoir, enhancing the downwards leakage gradient.

Diffuse leakage

Recharge to the Calivil Formation is generally thought to be by downwards leakage through the Shepparton Formation (Evans & Kellett 1989). Studies in northern Victoria have concluded that most leakage occurs where the Shepparton Formation is coarser grained close to the foothills, however Arad & Evans (1987) also found significant recharge downstream in the Campaspe valley.

Natural leakage was considered to be insignificant before development (Drury et al. 1984; Trewhella 1989). At any rate, diffuse leakage is likely to vary greatly due to heterogeneity of the subsurface.van der Lelij et al. (1987) estimated residual leakage to deep aquifers in the Coleambally area at 40–50 mm/year and local scale gross recharge rates at 120 mm/year. This contrasts with paddock scale estimates of gross recharge rates of up to 600 mm/year under rice in the Coleamballey irrigation area (Trewhella 1989).

In any case, diffuse recharge on the Lower Murrumbidgee alluvial fan is significantly greater than the average for the Murray Basin, which was estimated by Evans (1989) at about 1% of median annual rainfall. Evidence for dissipation of deep drainage recharge water to deeper aquifers is found in increasing groundwater levels in deeper aquifers.

Outflow

Outflows from the groundwater system include evaporation from shallow watertables, river base flows, playa lake discharge in the west of the basin, groundwater flow into the Murray Basin and groundwater pumping. Not all of these processes are readily quantifiable. Groundwater flow from the Murrumbidgee is a significant source for the Murray Basin at approximately 37 000 ML/year (Evans & Kellett 1989). Evaporative losses increase rapidly where the watertable is within 1.5 and 0.8 m of the surface, after which losses are relatively

Table 2.3: Recharge sources for the Lower Murrumbidgee

Recharge source	Method	Volume (ML/year)	Total recharge %
Total recharge		225 000	
Murrumbidgee River	modelling [1]	180 000	80
	gradients and transmissivities [2]	40 000	1% flow
diffuse leakage	modelling [1]	30 000–70 000	13–31
	rice water use @ 130 /year [3]	96 800	43
leakage to deep groundwater	downwards movement of saline gw [3]	<46 400	
	hydraulic grad/trans	30 000	
	best estimate	40 000	18
Storage	watertable rise	60 000	
channel leakage	420 km length, width 10 m [a]	6 000	2.7

1 Lawson 1996, 2 Woolley & Williams 1978, 3 Van der Lelij 1989
[a]K=5 mm/day for 10/12 mths

Table 2.4: Induced leakage at the Coleambally deep bore (Lawson et al. 1992).

Scenario	Downward flow velocity (mm/year)	Total pump test (ML)	K_v (m/day)
Non-pumping	7.3	2 500	10^{-4} to 10^{-2}
Pumping	13.0	1 030	

constant (van der Lelij 1990). Rice water use studies suggest losses via this process of up to 0.65 /ha/year.

Induced recharge

It was expected that agricultural development, including irrigation and deep groundwater pumping would result in greater recharge. This was in fact observed by Lawson & van der Lely (1992) who found that vertical leakage nearly doubled adjacent to a large volume irrigation bore (Table 2.4).

Induced leakage in the irrigation areas was expected to become an increasingly important source of recharge (van der Lelij et al. 1987). Recharge from the river has also been suggested as increasingly important:

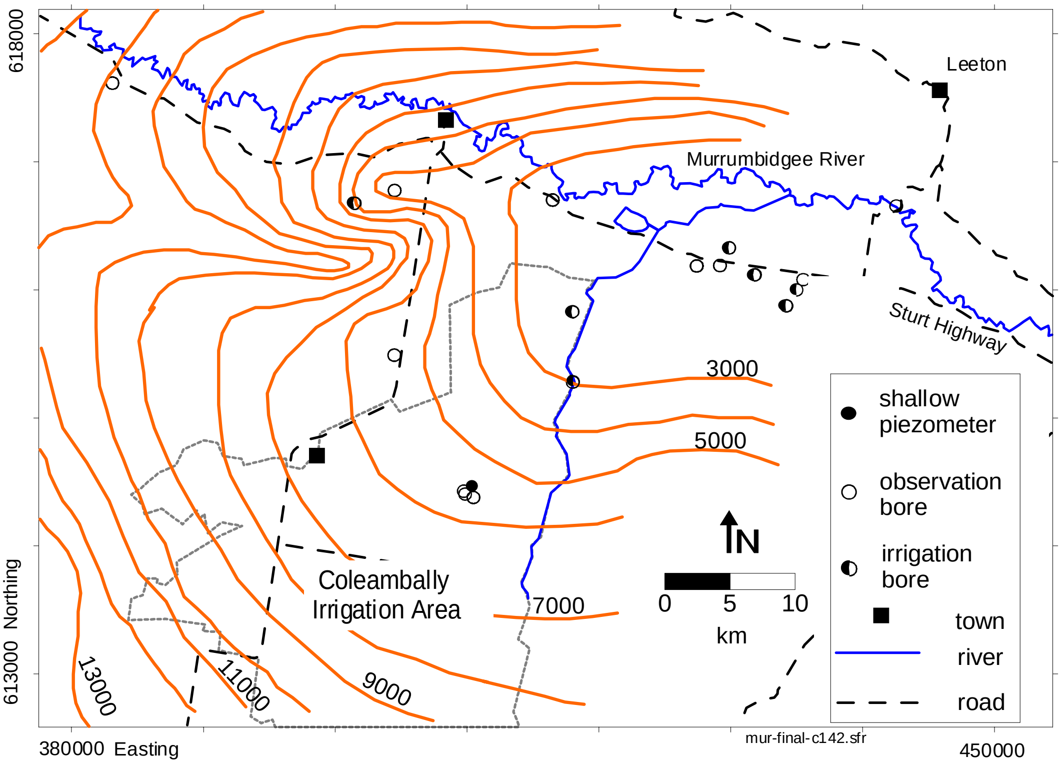


Figure 2.7: Groundwater residence times inferred from carbon-14. Contoured at 1000 year intervals. (after Drury et al. 1984).

> *" the amount of recharge which is currently occurring may have as much to do with the basin being "full" as to the physical ability of water to move from the main recharge source (the Murrumbidgee River) to the deep aquifers. It is possible that the most important change that may occur in recharge over time may actually be a response to groundwater extraction and its ability to facilitate additional recharge."* (Lawson & Webb 1998)

2.6.3 Implications of ^{14}C dating for recharge

Drury (1982) and Drury et al. (1984) reported on carbon-14 ages between modern and 15 800 years for groundwater in the Lower Murrumbidgee alluvial fan (Figure 2.7). Groundwater ages were greater in the west, and away from the highly transmissive sediments near the Murrumbidgee river. This data, indicate that the major recharge source is the river via the shoestring sands of the Shepparton formation. Dating anomalies in the area of greatest groundwater abstraction, were associated by Drury et al. (1984) with increased lateral flow and diffuse recharge through the overlying Shepparton Formation.

2.7 Groundwater quality

2.7.1 Review of hydrochemical studies

Various groundwater resources reports have referred to broad spatial trends of salinity and major ions (eg. Woolley & Williams 1978, Williams 1987, van der Lelij et al. 1987). However, to date there have been only two detailed hydrogeochemical studies of Lower Murrumbidgee groundwater, a pioneering effort by Kelly (1978), and a comprehensive hydrochemical anslysis of shallow groundwater by Williams & Ward (1987) and Williams et al. (1987). There has been no assessment of other water quality parameters such as nutrients or agrochemicals. Groundwater quality issues are increasingly on the agenda (MGMC), and the most recent groundwater status report recognised the need to incorporate groundwater quality information in the process of defining groundwater management zones (Lawson & Webb 1998).

Exploration of groundwater quality in the Lower Murrumbidgee began with Pels (1968) who sampled 5 windmills in the CIA area and analysed major ion composition, though the depth of groundwater intake was unknown. Kelly (1978) attempted the first comprehensive characterisation of hydrochemistry, sampling the Narrandera and Coleambally town water supply bores and approximately 20 bore sites along 6 sections in the mid and lower Murrumbidgee catchment. Unfortunately this data set is incomplete and is of limited use, lacking sampling dates, and in many cases, details of bore locations and depths. Increasing salinity was noted away from the Murrumbidgee River.

Williams (1987) reported that regional groundwater facies change from Na-HCO_3 type in recharge areas, through a transitional Na-HCO_3-Cl type to Na-Cl type in the discharge zone. Hydrochemical data from 8 bores around the CIA were presented, though the date of sampling was not specified. All groundwaters were found to be undersaturated with respect to calcite, gypsum and halite, except for some shallow groundwaters that were supersaturated with calcite. Of particular interest is the observation that shallow groundwater within the 20–30 m depth zone was highly variable.

A comprehensive statistical and mixing analysis of 178 shallow groundwater samples from selected areas of the CIA, MIA, and Darlington Point areas was presented by Williams & Ward (1987) and Williams et al. (1987). Examination of original data (B. Williams, pers. com., July 2001) revealed that these samples were obtained from the network of shallow piezometers rather than the DLWC observation bores that form part of this study. This study included a detailed analysis of the effects of mixing on SAR which was adjusted for calcite saturation. It was concluded that shallow groundwater may potentially supplement channel supply water with proportions varying from 25% in the saline areas to 95% in non saline areas.

2.7.2 Groundwater vulnerability

Piscopo (1997) rated most of the Lower Murrumbidgee alluvial fan groundwater as highly vulnerable, and a state wide aquifer risk assessment nominated it as the most highly vulnerable of all inland aquifers in New South Wales (DLWC 1998a). According to the Groundwater Protection Guidelines of the Agricultural and Resource Management Council of Australia and New Zealand (ARMCANZ 1995), a high vulnerability area requires a demonstrated remedial action plan prior to development. Approval of development depends on the capacity of the potential polluter to finance site investigations, on-going monitoring and an action plan for cleanup should the beneficial use be downgraded.

A groundwater vulnerability map of the Lower Murrumbidgee was prepared at a scale of 1:1 000 000 without local detail such as prior streams. The catchment response variable, which determined which areas were classified as vulnerable was defined as groundwater EC. For the Murrumbidgee, a median EC value of 1 480 μS/cm was used as broadly reflecting the division between short (lower EC) and longer (higher EC) residence times of groundwater. It was not possible to use depth to watertable (which may be a more appropriate vulnerability predictor) as a response variable due to the inaccuracy and scale of available data (G. Piscapo, pers.com., 30.5.01). This groundwater vulnerability map used the weights of evidence technique which included factors such as geology, minimum depth to groundwater and soil permeability. It was assumed that the vulnerable of groundwater in areas characterised by cracking clays was low.

Verification of the map was attempted with 30 samples analysed for nitrate which was found above 5 mg/L in low, medium and high vulnerability areas (unpublished DLWC data). No conclusive validation of groundwater vulnerability rating was possible using nitrates as an indicator. This failure was attributed to limited data size and bias in the selection of boreholes that were suitable for sampling.

2.7.3 Spatial variability of saline groundwater

The spatial distribution of shallow saline groundwater in and around the CIA was originally mapped by van der Lelij et al. in 1987. A comprehensive spatial analysis of groundwater salinity in each aquifer between the mid-1970s and 1987 was reported by Prasad et al.

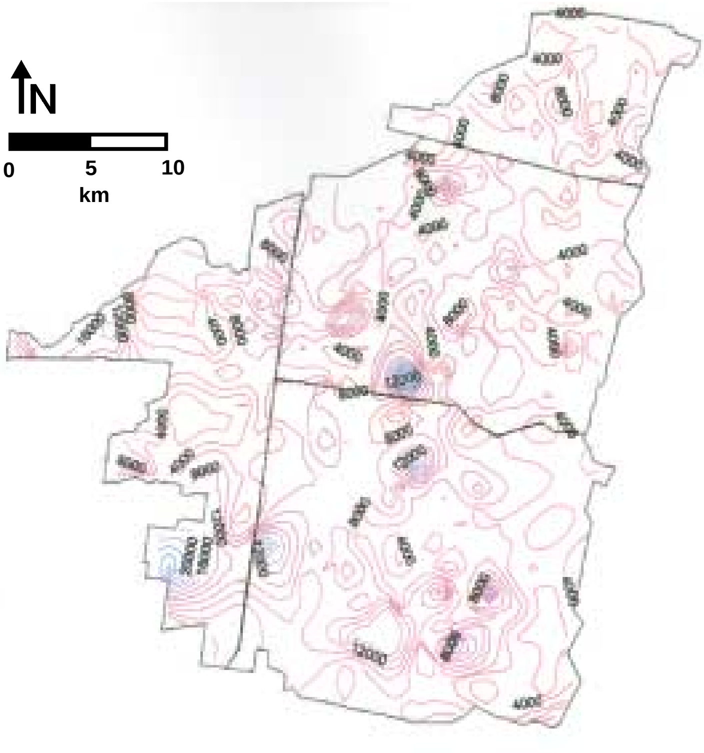


Figure 2.8: Spatial distribution of groundwater salinity, Shepparton Formation, September, 1999 (Christen et al., 2001).

(2001). Christen et al. (2001) presented a recent distribution of salinity in the combined Upper and Lower Shepparton Formation (Figure 2.8). Large variability was evident over short distances (eg. 8 000–12 000 μS/cm over a distance of 5 km), and very high salinity north-east of Coleambally and in the south-western CIA.

2.7.4 Observed and predicted water quality changes

The first observation of increasing groundwater salinity over time was noted by van der Lelij et al. (1987) who employed changes in vertical salinity profiles, together with water balance methods, to estimate the volume of groundwater dissipated to the deep aquifer in the CIA. The assessment was based on EC data from 2 sampling periods in 1966 and 1983 for piezometers between 15 and 35 m deep in the northern part of the CIA (Figure 2.9).

Lack of data limited analysis to averaged regional EC for 2 m depth intervals, rather than differences in individual piezometers. The resulting vertical EC profiles were then simplified to linear functions in order to estimate the depth at which EC was reduced

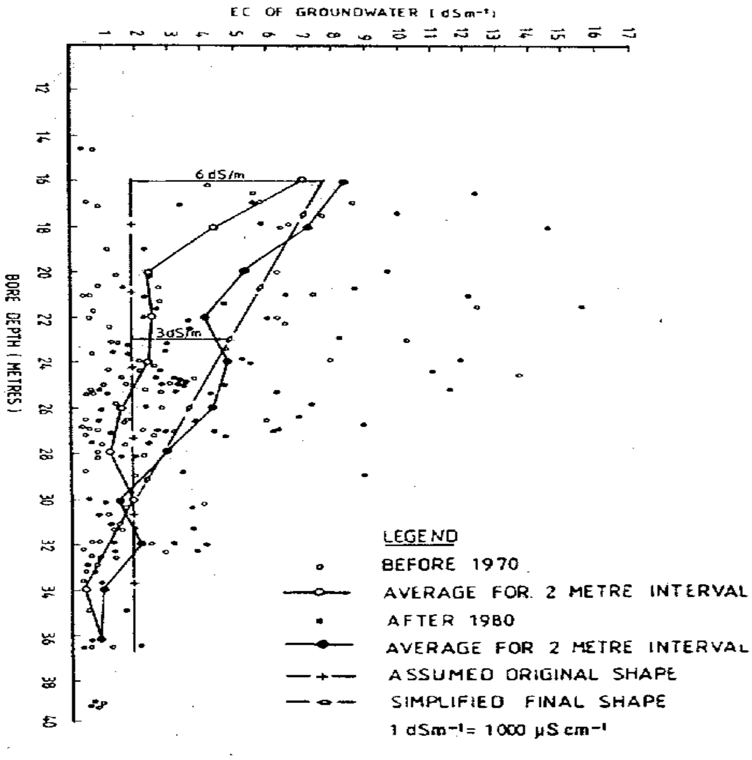


Figure 2.9: Observed changes in salinity with depth depicting downwards movement of salinity bulge between 1970 and 1980 (Van der Lely, 1989).

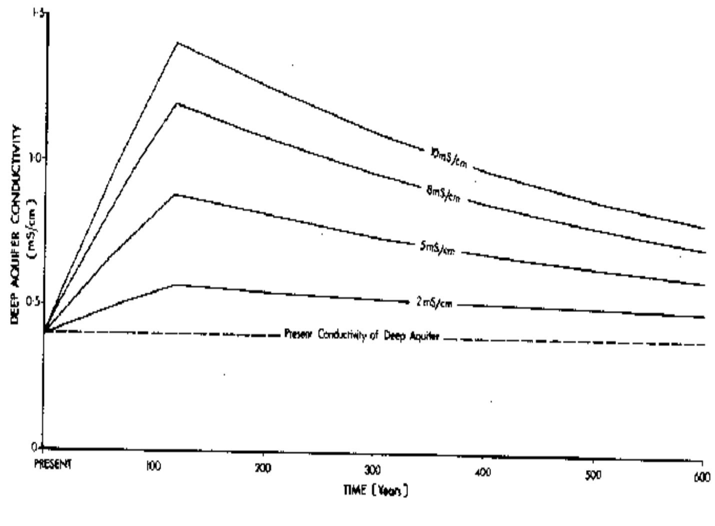


Figure 2.10: Predicted salinity of the deep pumped aquifers with natural leakage and full mixing of the shallow saline zone (Williams, 1987).

to half of the original value. Assuming that the half concentration point traveled at the mean Darcy velocity, assumed to be constant, groundwater was estimated to have moved downwards by about 5 m, equivalent to 10 ML/ha (assuming porosity of 0.2) over the 17 year period.

It was concluded that at this rate of leakage, the Calivil aquifer would be affected in 120 years and the overall risk of deep aquifer salinisation deemed to be small due to the relative volumes of saline and fresh groundwater.

Williams (1987) attempted to predict salinity changes in the Calivil aquifer that could be induced by pumping. The estimate was based on the distribution of shallow saline groundwater known at the time, (not including salt storage within clay aquitards) and assumed full mixing of saline groundwater of between 2 000 and 10 000 μS/cm with the large volume of groundwater stored within the Calivil aquifer (Figure 2.10).

Salinity was predicted to increase to a maximum of 3 000 μS/cm within the first 120 years, followed by dilution to background levels. Long term salinity increase of the deep aquifer was deemed to be insignificant. This study noted that the feasibility of deep aquifer pumping to mitigate shallow watertables depended not on water quality concerns, but on insufficient leakance data, and called for a full scale pump test which was to be subsequently carried out by Lawson & van der Lely (1992). The salinity of groundwater pumped from the CIA bore during testing increased from 510 to 670 μS/cm (Lawson & van der Lely 1992).

Prasad et al. (2001) found that salinity in the Upper and Lower Shepparton aquifer had increased over time (4 000–16 000 μS/cm) with a large area of saline groundwater. Curiously, salinity magnitude and distribution in 1987 was less than 1984. Groundwater salinity of the Calivil and Renmark aquifers were both less than 700 μS/cm and were similar in distribution, reflecting close hydraulic connections. There was no evidence for changing salinity at this depth.

2.8 Shallow saline groundwater

2.8.1 History and extent of problem

Severe water-table problems on a large scale first became apparent near Griffith in the 1930s during years of excessive rainfall (Hallows 1987). A committee set up in 1945 concluded that waterlogging was caused by deep drainage from rice fields located on permeable ground (Stannard 1978).

Originally, watertables in the Coleambally area were at about 20 m depth, with a slow rate of rise within the first 10 years of development and a rate of rise in mid 1980s of 1.6 m/year (Hallows 1987). This study assumed a porosity of 5% to calculate that 60 000 ML/year had been added to groundwater storage. At the time, it was thought that the CIA was to experience the largest rate of increase of any of the Riverine Irrigation areas, with a projected increase in shallow watertable areas from 1 000–45 000 ha between 1983 and 1995.

It was predicted that equilibrium would be attained by the end of the century if irrigation practices were to remain unchanged, although the lack of knowledge regarding watertable and salinity equilibrium was noted by Hallows (1987). Possible evidence that such an equilibrium was developing was observed in the gradual reduction of rice water use in the mid-1980s as watertables were higher and vertical pressure gradients had reduced (van der Lelij et al. 1987).

The delivery of irrigation water in a sustainable manner is now the responsibility of the landholder owned and operated, Coleambally Irrigation Corporation (CIC), subject to satisfying government licensing conditions. Groundwater levels are actively monitored and interpreted, based on 440 shallow piezometers monitored twice a year and 30 piezometers monitored bi-monthly (CIC 2000). The DLWC continues to monitor groundwater levels in the confined Calivil aquifer. Groundwater salinity is not regularly monitored, with the exception of a continuous EC gauge installed during construction of the CIA bore in the Calivil aquifer. The lack of monitoring is, in part, due to the narrow design of many shallow piezometers which prevents pump or bailer access.

The variability of area underlain by shallow watertables is depicted in Figure 2.11, relative to rainfall and the projected extent of the problem if no changes were implemented. Fortunately, scenarios suggested by Hallows (1987), and 'no plan' predictions of local water management plans, do not appear to have eventuated. Instead, a hyraulic equilibrium may have developed, possibly due to a combination of efforts to decrease recharge, and the watertable reaching a threshold depth at which capillary rise and evaporation balance accessions.

However, given the extreme variability from year to year in the area affected, due presumably to climatic variability, definitive trend analysis may be premature. The effect of climate on shallow watertable trends was explored in detail by Short & Khan (2000) and estimates of recharge based on water levels trends investigated by Christen et al. (2000), and will not be elaborated on here.

Recent soil sampling to a depth of 0.4 m, revealed that salinity has increased in the CIA, relative to surveys in the early 1990s (van der Lely 2000). This trend was attributed

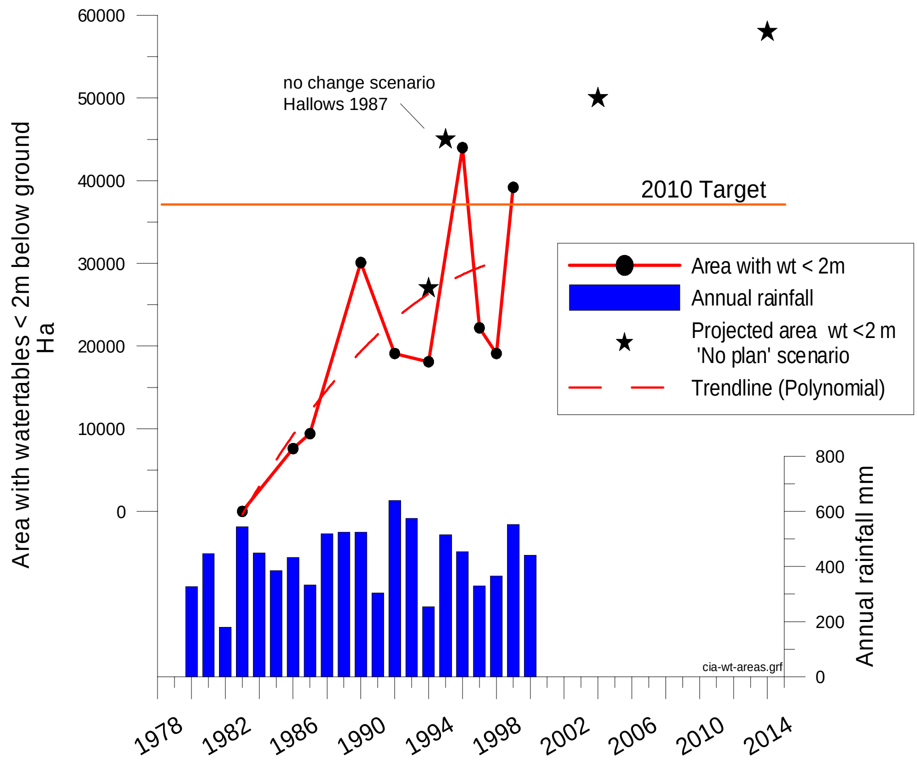


Figure 2.11: Actual and projected areas of the CIA with watertable depth <2 m from surface (Sources: Hallows 1987, CIC 2000, Bureau of Meterology data).

to a number of factors, but the highest correlation was observed with depth to watertable. This observation highlights the importance of watertable depth for irrigation sustainability.

2.8.2 Mitigation by groundwater pumping

One strategy to mitigate shallow groundwater has been, since the mid 1990s, to increase pumping of confined aquifers below the CIA. Two regional groundwater flow models, both based on the industry standard USGS MODFLOW package (McDonald & Harbaugh 1988), were developed to evaluate various shallow watertable management scenarios, including shallow and deep groundwater pumping.

The first of these was developed by DLWC using an element size of 56 km^2 with three layers - the Shepparton, Calivil and Renmark aquifers (Punthakey et al. 1994; Punthakey et al. 1995). Subsequently, a four layer model using an element size of 1.25 km^2 was developed by CSIRO Land and Water in Griffith, and continues to be refined in terms of hydraulic parameterisation (Pratha et al. 1997; Enever 1999; Khan et al. 1999; Prasad et al. 2001). The model incorporates hydraulic distinctions between the Upper and Lower Shepparton aquifers. Vertical leakances for each cell were determined from estimated vertical hydraulic conductivity and layer thicknesses indicated by borehole logs. The original model was unable to adequately represent seasonal groundwater levels, a deficiency attributed by Enever (1999) to high specific storage values for confined layers. This model has recently been extended (Prasad et al. 2001) to include the effects of pumping on groundwater salinity distribution by coupling MODFLOW with the solute transport simulator, MT3D (Zheng 1990).

Evans (1989) advocated greater usage of groundwater near the recharge areas in an effort to diminish the rate of groundwater rise elsewhere in the basin. Deep groundwater pumping subsequently became the favoured option across the Murray-Darling Basin with a number of schemes in northern Victoria (eg. SKM 1995; SKM 1997) and the installation of the CIA pump bore which commenced operations in November, 1990 (Lawson & van der Lely 1992).

An analysis of induced vertical leakage through the Shepparton Formation was based on a 468 day pump test of the CIA pump bore during the initial experimental phase. Relative to control areas, the shallow watertable was lowered 0.5 m within 2 km of the pumping bore and a head difference of about 10 m developed between the Shepparton and Calivil formations. An effective hydraulic barrier was formed by low permeability clay layers at depths of 25–50 m.

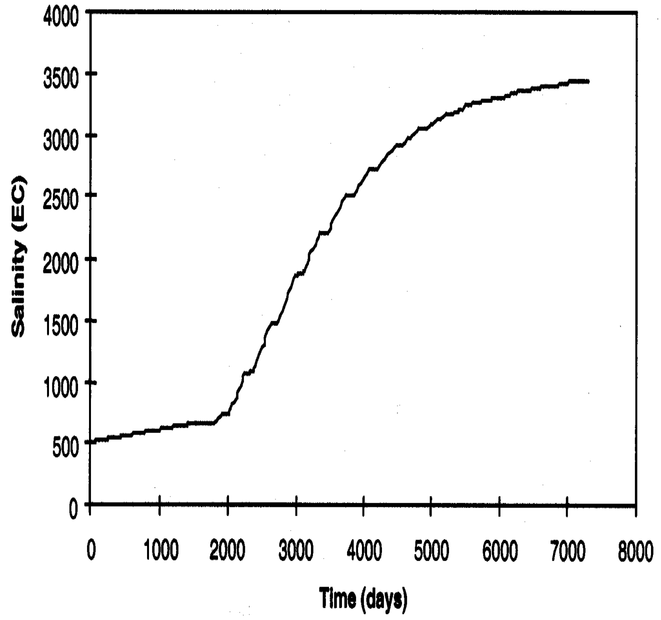
The deep groundwater pumping scenario considered by Punthakey et al. (1994) was based on two pumping bores located in areas of the CIA likely to be most seriously affected by shallow watertables, and where groundwater salinity of the Calivil aquifer was less than 1000 mg/L. Decreased groundwater levels resulted in large areas of the Calivil aquifer, and to a lesser extent in the overlying Shepparton and underlying Renmark aquifers. It was concluded that deep groundwater pumping was the best long term option for controlling high water tables but warned:

> *"a major study of deep groundwater pumping would be required to investigate the long term impact of water quality of the Calivil, the maximum permissable drawdown that will be required, and the extent of mixing of groundwater from the Shepparton and Renmark aquifers"*

Lawson & Holland (1999) reviewed the role of deep groundwater pumping in watertable control, concluding that it was not an effective watertable control option because of poor hydraulic connections and potential leakage of shallow saline groundwater. A number of pump tests on the Riverine Plain were compared with findings from the Indus Basin in Pakistan where the ratio of aquifer K_h to aquitard K_v was much lower and watertable control pumping schemes were more successful.

Recently, five alternative pumping strategies were evaluated by Prasad et al. (2001), using the more refined regional flow model previously described. These strategies considered one or two bores at three potential locations, and the proportion of groundwater pumped from the Calivil versus the Renmark aquifers. The greatest area of drawdown within the CIA that corresponded with the smallest aquifer salinity increase, was found to be achieved by pumping at two locations (southern central CIA and western CIA). A total abstraction of 4 000 ML from these locations was modelled, with 30% from the Calivil and 70% from the Renmark aquifer. This abstraction resulted in increased salinity of the deep aquifer from 600 to 1 000 μS/cm over 15 years (Figure 2.12). Groundwater salinity impacts could be minimised by locating bores in areas with poor vertical connection, and by pumping a greater volume from the Renmark than the Calivil aquifer. Existing leakage to deep aquifers was increased by only 10% by this scenario.

It was concluded that strategic pumping of deep aquifers could complement other measures such as shallow groundwater pumping, by stabilising and preventing future problems. Although deep groundwater pumping would potentially affect large areas, benefits would not become apparent for many years and would vary spatially depending on aquifer connections.



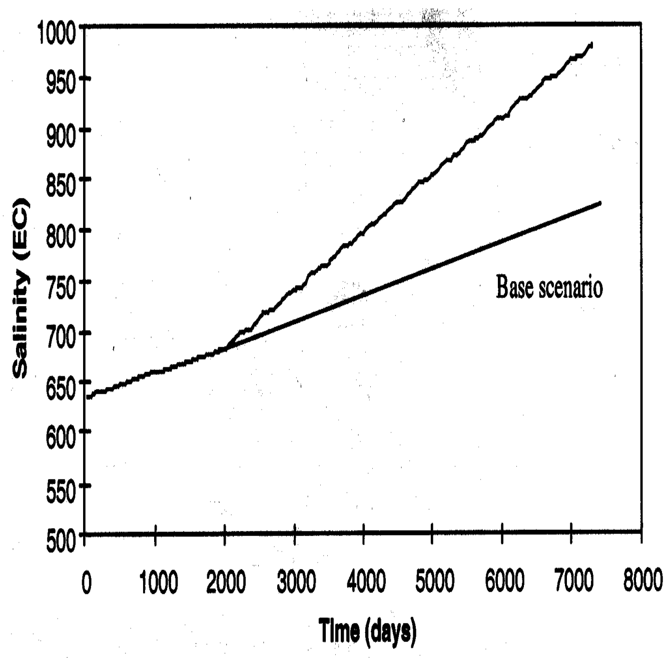


Figure 2.12: Projected salinity increase of pumped groundwater over time. **A** Worst case scenario - a single bore pumping entirely from the Calivil aquifer and **B** Recommended best case scenario - one of two strategically located bores pumping 30% from the Calivil aquifer and 70% from the Renmark aquifer (Prasad et al. 2001).

2.9 Summary

- Of the major aquitard deposits in the Murray-Darling Basin, only the deep Tertiary age Geera clay has been studied in detail. There is extensive literature describing soils, paeolodrainage and aeolian deposits in the Lower Murrumbidgee area, but these are limited either to surface outcrop, or shallow profiles.

- Groundwater is pumped for irrigation purposes (and some town supplies) predominately from the confined Calivil and Renmark aquifers, with increased pumping since the mid 1990s to mitigate shallow watertables. The shallow Shepparton aquifer, which is saline in some areas, is used mainly for stock water. The groundwater system is over allocated, with long term declining groundwater levels in some areas, but increasing groundwater levels in other areas.

- The only detailed hydrochemical study available focused on shallow groundwater, which is vulnerable to groundwater quality decline. A bulge of saline groundwater was found to have migrated downwards, but predictions suggested that the salinity of the Calivil aquifer would increase only by a small amount over several hundred years.

- Carbon-14 dating of the Lower Shepparton and Calivil aquifers indicated major recharge from the Murrumbidgee River as it discharges onto the Riverine Plain, although there was uncertainty in actual aquifer residence time which were double hydraulic estimates.

Chapter 3

LITERATURE REVIEW

In this chapter the current state of knowledge pertaining to hydrogeological and hydrogeochemical processes within aquitards, and between aquitards and adjacent aquifers is introduced. Insights from many international aquitard studies are discussed, with a bias to North America and Europe, where a great many advances have been made. Particular focus is devoted to issues that are pertinent to the present study of clayey aquitards.

3.1 The nature of aquitards

The majority of hydrogeological investigations focus on the development of aquifers, whilst virtually ignoring the role of aquitards in groundwater flow and contamination. However, Back (1986) points out, that even in the late 1880s and early 1900s, groundwater development relied on natural pressures of artesian aquifers, and studies at the time recognised the critical role of confining layers and formulated key hydrogeological concepts. With the development of the turbine pump and drilling technologies, development of high yielding deep aquifers surged, though workers remained somewhat oblivious to intervening low permeability units. Knowledge of aquitards has thus lagged behind our understanding of aquifers.

Aquitards are, in geological terms, argillaceous sediments which are fine grained. Fine grained detrital sediments and rocks are the most abundant on earth, forming 50–75% of the geological column (Stow 1981).

Aquifer and aquitard are relative terms such that a silt may be an aquifer in a clayey environment, but form an aquitard in sandy sediments. For example, early studies in the Murrumbidgee area by Talsma & Flint (1958) defined an impermeable layer as having ≤ 0.1 the hydraulic conductivity of the overlying layer. Neuzil (1986) noted that no properly tested geological media were totally impermeable. These terms can apply to

Table 3.1: Distinguishing aquitards and aquifers

Aquitards	Aquifers
Fine-grained	Coarse-grained or fractured
Dominated by reactive minerals	Quartz dominated
Vertical flow dominates	Horizontal flow dominates
High porosity & storage	Medium porosity & storage
Low permeability $<10^{-2}$ darcy	High permeability $>10^{-2}$ darcy

specific sedimentary layers or complete geologic formations depending on the context.

Common charactersitics that distinguish aquitards from aquifers are summarised in Table 3.1. Back (1985, 1986) considered the diverse roles of aquitards in flow regimes, as indicated by the following list:

- Influence topography and drainage patterns.
- Cause swamps and perched watertables.
- Influence the relationship between water table and the potentiometric surface of confined aquifers.
- Constrain the flowpath.
- Control the rate of infiltration and discharge.
- Prevent CO_2 degassing.
- A source and sink for leachable chemicals.
- Serve as a natural filter for attenuation of chemicals.
- Provide ion exchange sites.
- Cause fractionation of isotopes.
- Source of exploitable water.

In Australia, aquitard seepage was recognized by Ramamurthy & Holmes (1983) as major limitation to understanding and modeling of groundwater flow due to lack of information regarding thickness, extent and hydraulic characteristics of aquitards. Complexity of aquitard physical and hydraulic characteristics were considered to be greater for aquitards than for aquifers.

3.1.1 Characteristics of clay minerals

Clay minerals are generally the dominate consituent of aquitards and are critical controls of permeability and reactivity. Clay minerals are sheet silicates comprised of silica tetrahedron sheets and sheets of hydroxyl octahedron with Al^{3+} or Mg^{2+} at the centre. A net

Table 3.2: Characteristics of common clay minerals (Rowe et al. 1995)

Characteristic	Montmorillonite (bentonite)	Illite (mica)	Kaolinite
Structure	2:1 (3 layers)	2:1	1:1 (2 layers)
CEC (meq/100g)	100	25	5 (little electrical charge)
Surface area (m^2/g)	800	80	15 (large crystals)
K (m/s)	10^{-11} to 10^{-15}	10^{-9} to 10^{-10}	10^{-8}
Activity	highly active with c-axis swelling from 1.7 to 10 nm depending on chemical state	normal activity, efficiently compacted	inactive, not well compacted

negative charge is balanced by cations which occupy inter-sheet sites, where the cation exchange capacity (CEC) is defined by the number of cations required to neutralize 100 g of clay.

Here, characteristics of clay minerals that are of particular importance to groundwater systems are briefly reviewed with reference to aquitard investigations. An introduction to clay genesis and mineralogy is provided by other sources including Velde (1995), Chamley (1989), and Wilson (1999). It is sufficient here to note that clay genesis depends on whether the environment is alkaline or acidic, and on the type of source minerals present and consequent solute concentrations of Si, Al, K and Fe. The degree of leaching which reflects climatic and topographic factors is also important.

The characteristics of common clay minerals are summarised in Table 3.2. Montomorillonitic clays such as bentonite and smectite have a large negative charge as indicated by large CEC, and are small platelet structures with very large surface area. By contrast, kaolinite has little electrical charge, and large crystals, and so has a small CEC and surface area. The reactivity, and permeability of clay reflects these physical properties.

Formation of authigenic clay can be an important control on groundwater flow since illite forms hairy aggregates that clog pore throats and kaolinite forms neat booklets that may also reduce permeability (Appelo & Postma 1996).

Active smectite clays dominate the vertisol soil group of black self-mulching cracking clays (Ug soils) (Northcote et al. 1975). These soils form in regions of 500–1 000 mm of annual rainfall, mainly on undulating plains with a source of basic parent materials and are common in Africa and India. In Australia such clays characterise 48 Ma ha and are common in prime agricultural areas (Hubble 1981). Moist to wet soil has a very low permeability and a high run-off potential, with the lack of leaching reflected by the presence of carbonates and soluble salts in the deeper subsoils.

3.1.2 Permeability

It may be assumed that permeability is equivalent to the intrinsic permeability of geological material, since there is little variation in fluid density and viscosity in shallow groundwater systems (van der Kamp 2001). In subsequent discussion the terms permeability and hydraulic conductivity, K (dimensions L/T) are used interchangeably.

Low permeability material is defined by Neuzil (1986) as K $<10^{-8}$ m/s, although permeabilities much lower than this have been recorded (eg. 10^{-16} m/s). By reviewing a number of data sets, Neuzil (1986) established a log-linear relationship between permeability and porosity, with an order of magnitude decrease in permeability for each 0.13 decrease in porosity. Further, it was found that permeability commonly varied over 3 orders of magnitude for similar porosity (Neuzil 1994), a range unusual among natural parameters.

On this basis, it is reasonable to seek order of magnitude permeability measurement by field, laboratory technique, or inverse flow modelling techniques. van der Kamp (2001) reviewed methods to determine in situ hydraulic conductivity of aquitards including: slug tests, pumping tests, response to mechanical loading and analysis of natural pore pressure fluctuation. It was recommended that field based methods resulted in more reliable permeability values than laboratory values, at least for the purpose of regional scale leakage analysis.

Permeability may be scale dependent or independent (Neuzil 1994; van der Kamp 2001). Scale independence for glacial till has been demonstrated by Keller et al. (1989) where laboratory values agreed with inverse estimates at scale of 10s of meters, and by Hendry & Wassenaar (1999) who compared groundwater velocity derived from diffusion profiles of δ^2H with hydraulic testing. By contrast, other studies of shallow clay aquitards have found permeability to increase with scale of investigation (eg. Keller, Van der Kamp, & Cherry 1988, Rudolph et al. 1991, Gerber & Howard 2000). Scale dependency is usually attributed to inferred heterogeneity that exist at field scale either via fracture or higher permeability aquitard windows.

Gerber & Howard (2000) studied vertical leakage through a regional till aquitard in southern Ontario. The permeability of core samples was determined at 10^{-11} to 10^{-10} m/s. However tritiated water was found below the till, suggesting preferrential flow along sand lenses, erosional surface, joint and fractures in the matrix. Slug test, pump test and water balance studies resulted in permeability estimates from 10^{-12} to 10^{-5} m/s. A numerical flow model was calibrated to waterlevels and discharge measured by varying the regional bulk K. Best fits were obtained with 5×10^{-10} to 5×10^{-9} m/s, with regional recharge between 30–35 mm/year.

3.1.3 Specific storage

Specific storage is the amount of water per unit volume of a saturated formation that is stored or expelled due to compression of the matrix and pore water, per unit change in head (Fetter 1994).

Historically, aquitards were assumed to play a secondary and passive role in transmitting water to confined aquifers. Neuman & Witherspoon (1969b) sketched the development of methods to account for leakage that began with the work of DeGlee (1930) and Jacob (1946) who both assumed that leakage was proportional to the change in hydraulic head across the aquitard, ignoring storage capacity. The widely used Hantush & Jacob (1955) solution for leaky aquifers continued this line of reasoning. The contibutions of aquitard storage were first accounted for with the modification of Hantush (1960).

Neuman & Witherspoon (1969a, 1969b) discussed potential errors of early methods, noting a general lack of data on hydrogeological characteristics of aquitards. Available specific storage values for aquitards suggested that storage was equal and possibly greater than confined aquifers and could not be neglected. A solution for quantifying flow to confined aquifers that included this source was subsequently developed (Neuman & Witherspoon 1972).

Back (1985) reported that most of the water released from storage during development of the Dakota aquifer originated from bounding aquitards. Aquitards themselves could therefore constitute a significant source of groundwater when depressurised. Dewatering of aquitards accounted for up to 40% of groundwater pumped in the San Joaquin valley (Schmidt 1977).

Field tests that include piezometers within the aquitard, are the preferred method for determining specific storage (van der Kamp 2001). A review of a number of studies, found that values obtained by pump tests and derived from barometric response were typically 1–3 orders of magnitude smaller than laboratory consolidation tests.

3.1.4 Pore pressure response

Pore pressure changes within clay aquitards can be significant due to compressibility of the matrix. Compressibility of the aquitard matrix (or aquifer skeleton) is related to S_s by Equation 3.1 (Fetter 1994).

$$S_s = \rho_w g(\alpha + \phi\beta) \tag{3.1}$$

Where S_s is specific storage (1/m), ρ_w is density of water (kg/m^3), g is acceleration of gravity (m/s^2), ϕ is porosity, α is compressibility of aquitard matrix $(\frac{1}{N/m^2})$, and β is compressibility of water $(\frac{1}{N/m^2})$.

The implications of compressibility for groundwater level interpretation is not commonly considered in groundwater resource investigations and is covered in detail by only one major groundwater text, Domenico & Schwartz (1990). Early advances were made within the disciplines of geotechnical engineering and soil mechanics including the classical work of Terzaghi (1923) and Biot (1941).

Terzaghi recognized that explusion of pore fluid during consolidation proceeded at a rate limited by hydraulic dissipation. In Terzaghi's 1D consolidation equation (Equation 3.2), compressibility of water is assumed to be negligible relative to pore compressibility (Domenico & Schwartz 1990). It is analogous to Fick's 1st law which describes flux of a chemical due to a concentration gradient, based on a diffusion coefficient.

$$\frac{\partial P_{ex}}{\partial t} = c_v \frac{\partial^2 P_{ex}}{\partial z^2} \tag{3.2}$$

Where P_{ex} is transient pore water pressure in excess of hydrostatic value, c_v is coefficient of consolidation , z is depth, t is time.

The rate limiting term c_v, or coefficient of consolidation, is equivalent to the parameter hydraulic diffusivity, h_{diff}, in groundwater hydrology (Domenico & Schwartz 1990) and is related to K and S_s by Equation 3.3.

$$c_v = h_{diff} = \frac{K}{S_s} \tag{3.3}$$

Where h_{diff} is hydraulic diffusivity, K is hydraulic conductivity and Ss is specific storage.

The complexities of interpreting groundwater levels in low permeability environments has been extensively explored (van der Kamp & Gale 1983, Rudolph & Frind 1991, Neuman & Gardner 1989, Neuman et al. 1982, van der Kamp & Maathuis 1991, Rojstaczer 1988 and Sepulveda & Zack 1991). In North America, variable water levels have frequently been interpreted in the light of the relationship between matrix compressibility and changes in effective stress (Landmeyer 1996, Hare 1997, Rasmussen & Crawford 1997, van der Kamp & Maathuis 1991). Application of these principles to confined aquifers in New Zealand has been reported by Bardsley & Campbell (1999) and Bardsley & Campbell (1994). Shallow groundwater level variability within smectite clays on the Liverpool Plains, has recently been reinterpreted as a response to loading events (Timms et al. 2001; Timms et al. 2002).

Rapid water level changes observed in piezometers which are screened within clayey sediments (or even sands) can easily be misinterpreted to be a consequence of significant groundwater flux due to recharge or leakage, even though the units are practically impermeable.

3.2 Leakage processes

Groundwater leakage to confined aquifers may originate either from within the aquitard (specific storage) or from overlying aquifers. Flux from overlying water sources may pass through the aquitard, but will preferentially bypass aquitards if fractures or other higher permeability conduits are present. In any case, aquitard permeability has been identified as the most significant uncertainty in quantifying subsurface flow (Neuzil 1994).

The hydrogeological term for the 1-dimensional leakage is Darcy flux, which is definined as the apparent velocity of groundwater as if in an open conduit (Equation 3.4). This parameter is distinct to the actual linear velocity which is found by dividing the Darcy flux by effective porosity.

$$V_a = -K\frac{dh}{dL} = \phi_e V \tag{3.4}$$

Where V_a is Darcy flux, ϕ_e is effective porosity and V is average linear velocity.

3.2.1 Aquifer-aquitard permebility contrasts

Vertical leakage is limited by the bed of lowest permeability, no matter how thin. Flow lines are refracted when entering a matrix with contrasting hydraulic conductivity. The greater the contrast the greater the angle of refraction (Fetter 1994). In many groundwater systems, it is the permeability contrast between aquitards and aquifers that determines leakage (Table 3.3).

Regional differences in aquifer K and aquitard K are particularly important in large sedimentary basins where topographical gradients results in a small driving force (Belitz & Bredehoeft 1990).

Table 3.3: Relative aquifer and aquitard K, selected examples

Location	aquifer K_h (m/day)	aquitard K_v (m/day)	ratio K_h:K_v
Indus Basin, PAKISTAN [1]	50.0	1-10	50:1 to 5:1
Denver Basin, USA [2]	0.0023	<0.000 000 23	>1 000 000:1
South Dakota, USA [2]	0.015	0.000 002 3	<1 000 000:1
Bighorn Basin, USA [2]	0.023	0.000 000 69	<1 000:1
CIA pump bore, AUST [3]	50.0	0.000 1	500 000:1
Murray Valley, Vic., AUST [4]	5.0	0.000 6–0.000 42	83 000:1 to 11 900:1
Campaspe West, Vic., AUST [4]	40.0	0.001 to 0.01	40 000:1 to 4 000:1
Campaspe, Vic. AUST [6]	15-65.0	0.004	3 750:1 to 16 250:1

1 Mundorff et al. 1976, 2 Belitz & Bredehoeft 1990, 3 Lawson & van der Lely 1992, 4 SKM 1997, 5 Brownbill 1997, 6 Lawson & Holland 1999

The importance of relative permeability between a shale aquitard and sandstone aquifer in controlling vertical recharge in northwest New Mexico, was investigated by Stephans (1983). Recharge to the dipping aquifer was thought to predominately occur at outcrops with groundwater flow downdip within the sandstone aquifer. However, a review of available water level and permeability data led to a model that simulated two scenarios of relative permeability. The relative permeability for shale aquitard:alluvial cover:sandstone aquifer was 1:10:100 for case 1 and 1:1 000:100 000 for case 2. The first case resulted in a flow divide within the dipping aquifer, such that most recharge was by vertical leakage through the shale aquitard causing a flow divide and flow upwards along dip toward outcrop areas. In case 2, a 100 000 factor permeability contrast between the aquifer and aquitard, resulted in most recharge occurring at the outcrop area.

Though this example deals with large scale semi-consolidated units with significant structural elements, similar principles may be applied in unconsolidated sediments where bedding contacts are sub-horizontal.

3.2.2 Macropore flow

Rapid bypass flow may occur through macropores such as fractures, animal burrows and root tubes (Wood et al. 1997). Bore casing or annulus leakage may be considered an anthropogenic form of macropore flow, the effects of which will be considered in Chapter 3.5.4.

Microfractures may not be visible in cores (eg. Keller et al. 1986), but may be revealed by X-ray radiographs (eg. Doser et al. 1998) and deep fractures have been observed to penetrate far below surficial weathered zones (D'Astous et al. 1989).

Table 3.4: Examples of aquitard effects on regional groundwater resources

Location	Nature of aquitard	Effects on underlying aquifer
Dakota aquifer, USA [1]	confining shale	vertical leakage predominate source of recharge prior to development, most water released from storage of aquitards
NW New Mexico, USA [2]	dipping, layered shale aquitard & sandstone aquifer	relative differences in aquitard/aquifer permeability control flow directions indicate major aquitard leakage recharging many areas
Southern Ontario, CAN [4]	thick deposits of glacial and interstadial alluvial till	vertical leakage primary source of recharge for domestic and municipal water supply wells
Liverpool Plains, AUST [5]	fractured, semi-saturated smectite clay	permits leakage of irrigation drainage water to the shallow aquifer

1 Bredehoeft et al. 1983, 2 Stephens 1983, 4 Gerber & Howard 2000, 5 Timms & Acworth, 2002

Streltsova (1976) proposed a classification scheme for fractured medium according to permeability and porosity of blocks or peds relative to fissues. Purely fractured medium (eg. rock) consists of continuous fracture porosity. In a double-porosity medium hydraulic properties of block and facture were of the same order of magnitude, with permeability controlled by fracturing. A heterogeneous medium was distinguished from a double-porosity medium by infilled fractures (eg. with silty clay or fine sand).

The effective hydraulic conductivity, K_e, of fractured deposits may be several orders of magnitude greater than unfractured medium. Snow (1969) showed that K_e is sensitive to fracture aperature, and less sensitive to fracture spacing.

$$K_e = \left(\frac{\rho g}{\mu}\right)\left(\frac{N(2b)^3}{12}\right) \tag{3.5}$$

Where ρ is fluid density, μ is dynamic viscosity, b is fracture aperature and N is the number of fractures per unit length of medium.

This means that small fractures (eg. 30 μm) may significantly increase K, but impart a negligible effect on bulk porosity or bulk storage. Harrison et al. (1992) found that 80% of total flow through an aquitard could be attributed to fracture flow, even when fracture porosity (fracture volume:bulk volume) was only 1.3×10^{-5}. In a semi-arid area of Texas and New Mexico, macropore flow accounted for 60–80% annual average recharge of 11 mm (Wood et al. 1997).

Fracture flow is commonly identified by scale independence of hydraulic conductivity determined by different methods. For example, Rudolph et al. (1991) calculated K_e of 1.5×10^{-8} m/s from depth profiles of major ions, 2.7 times higher than K obtained by consolidation testing of 5.5×10^{-9} m/s. In glacial till K field hydraulic conductivity was up to 3 orders of magnitude greater than K of 2×10^{-10} m/s determined by oedometer tests for the clay matrix (McKay et al. 1993). Keller et al. (1988) documented bulk permeability 2 orders of magnitude greater than matrix permeability for a clayey till near Saskatoon. Recent investigations at an irrigated site near Gunnedah, Liverpool Plains (Timms & Acworth 2002a) found evidence for irrigation drainage leaking through 15 m of fractured semi-saturated silty-clay. Soon after flood irrigation began, the salinity of the shallow aquifer decreased from 1080 to 365 mg/L.

3.2.3 Leakage through aquitard windows

Leakage through zones of higher permeability, or aquitard windows, have been neglected during groundwater resource investigations (Gerber et al. 2001) with a handful of notable exceptions. The deficit of literature on this topic is contrary to the likely prevalence of aquitard windows in many groundwater environments. This may be attributed principally to the intermediate scale of the problem, which falls between small, laboratory testable features such as fractures and large scale regional problems that are described by flow models. The difficulty of defining subsurface stratigraphy with a limited density of boreholes and other geophysical information, further compounds the problem.

Aquitard windows may be classified as a form of macroscopic heterogeneity, according to five scales of heterogeneity which were outlined by Galloway & Hobday (1996). These aquitard windows (scale of tens to hundreds of meters), reflect sedimentary facies relationships. Relationships between individual deposits must be accurately delineated if deterministic simulation of fluid flow is to be achieved.

Theoretical developments were made by Strack (1981a, 1981b) who developed approximate solutions for two-dimensional flow in aquifers with clay laminae and modelling of theoretical groundwater systems with windows by Strack & Haitjema (1981a, 1981b). The analysis assumed that the clay laminae were infinitely thin and impervious and did not overlap (but could be at different elevations), and was limited to steady state fresh water flow. It adopted the Dupuit-Forchheimer approximation that reduced three-dimensional flow to two-dimensional horizontal flow. Flow net analysis demonstrated that the volume of water flowing through an opening was insensitive to the window width of a length at least 0.2% of aquifer thickness.

A recent study by Gerber et al. (2001), presented field evidence for leakage pathways along vertical sedimentary structures such as sand dikes and dipping erosion surfaces within glacial till in Ontario, Canada. Discontinuous sand and gravel lenses were also implicated in lateral groundwater transmission to zones of fractured till whereapon further vertical leakage occurred. Such step wise vertical and leakage pathways explained very rapid rates (>1 m/year) of flow within a thick till with low permeability matrix. An assessment of the importance of large scale heteorogeneity was limited however because of the small number of outcrops and test pits. By contrast, quantitative assesement of the role of fracturing was achieved through field and modelling techniques.

A numerical model for this same area of glacial till was developed which assessed sensitivity to aquitard window features (Martin & Frind 1998). The model was based on a comprehensive analysis of available hydrogeological data. It was concluded that aquifers were recharged predominately by flow through high permeability windows. Well capture zones were found to be highly sensitive to geologic structure, though water table levels were not significantly affected.

The application of geostatistical mapping techniques to calculate the probability of aquitard windows was demonstrated by Desbarats et al. (2001). The study concluded that heterogeneities associated with aquitard windows have the largest effect on regional groundwater flow patterns. A map using a $250\,m^2$ grid spacing was produced depicting the probability of encountering an aquitard window. The spatial trends and alignments of probable aquitard windows reflected the courses of incised channels. Significant inter-aquifer flow was thought to occur through these windows, although leakage through the till matrix was also important on a local scale.

The dangers of assuming laterally extensive aquitard layers is illustrated by a case study in Siberia which is highlighted by MSU (2000). Regional flow modelling, based on existing conceptual models of the local groundwater system, had predicted that liquid containing radioactive species could be safely disposed of below an aquitard. The subsequent contamination of groundwater in an adjacent aquifer was attributed to upwards leakage through aquitard windows, the existence of which had been previously unknown.

Aquitard windows in alluvial fans and fluvial deposits

Landscape holes have been virtually neglected in alluvial fans and fluvial deposits (Williams et al. 2001), even though such leakage pathways may be of greater importance than micro-pore and macro-pore heterogeneity.

A practical example of groundwater flow analysis in a complex alluvial aquifer system containing interbedded silts and clays was provided by Fogg (1986). The importance of interconnected flow pathways over the permeability of individual deposits was shown by adopting an integrated geologic-hydrological approach for the Wilcox aquifer system in Texas. Hydraulic head was not sensitive to the degree of hydraulic connection, in the numerical flow model for this system, showing only a subtle response to paleochannel flowpaths where large velocity vectors occurred. Since these purtubations were within the measurement error for hydraulic head, a calibrated model that reproduced hydraulic head could fail to adequately depict leakage which was of significance for solute transport. Uncertainties regarding aquifer-aquitard architecture and interconnectiveness led to a conclusion that:

> *"...we are only slightly better prepared to model flow and solute transport realistically in many types of porous media than in fractured media."*

3.3 Hydrogeochemical role of aquitards

The geochemical influence of aquitards on groundwater systems has been thoroughly reviewed by Back (1986). Their hydrogeochemical dominance in the subsurface is due to reactive minerals that are soluble or have high ion-exchange capacity.

The geochemical role of shallow aquitards has been generally overlooked, as the oil and gas industry focused attention on very deep aquitard systems. While many principles have since been adapted and applied to groundwater studies, there are differences. Shallow confining beds are generally more inert than deeper ones, due to leaching of soluble minerals. Also, membrane filtration (eg. ultrafiltration that is pressure induced flow and osmosis where water flows towards a high ionic strength solution) processes which control deep brine evolution are not as important in shallow systems due to rapid flow paths and the lack of concentration contrasts (Back 1986).

3.3.1 Matrix interactions

One pertinent example of matrix interactions is the application of ^{14}C dating within low permeability units. Underestimation of age occurs where unknown losses of ^{14}C occur in

Table 3.5: Importance of factors that result in under or overestimation of ^{14}C groundwater dating
.

Effect on estimated age	Process
Under	Matrix diffusion into dead-end pores or micro-fractures [1]
Under	Matrix diffusion into bounding aquitards [2]
Under	Diffusion of CO_2 through unsaturated zone to unconfined aquifers [3]
Over	Dissolution of $CaCO_3$ [3]
Over	Sulphate reduction and organic matter oxidation [3]
Over	Mantle sources of CO_2 to deep groundwater [3]
Over	Methanogenesis $2CH_20 \Rightarrow CO_2 + CH_4$ [3]

1 Sudicky & Frind 1981, Maloszewksi & Zuber 1991, 2 Sudicky & Frind 1981, Sanford 1997, 3 Walker & Cook 1991, 4 Clark & Fritz 1997

the system, while overestimation of age is caused by dilution of ^{14}C by old 'dead' carbon (Table 3.5).

Underestimation of groundwater residence time in clay may be attributed to matrix diffusion into dead-end pores of the low permeability formation (Maloszewski & Zuber 1991; Sudicky & Frind 1981). It is noted by these authors that this affect occurs to a greater extent the longer the residence time, thus compounding the errors.

Simple correction equations, such as those of Tamers (1975) and Pearson (1965) may be used to improve estimates of residence time, depending on specific geochemical system and data available (Clark & Fritz 1997). However, Hendry & Schwartz (1988) cautioned against applying such simplistic approaches in lieu of numerical models that are better able to account for complex hydrogeochemical processes that occur within groundwater systems. Furthermore, ^{14}C age correction equations, are not appropriate for clayey material where diffusion is the dominant transport mechanism (Wassenaar & Hendry 2000). These models assume piston flow that ignores dispersion and diffusion, resulting in overestimation of residence time for aquitard pore waters.

3.3.2 Solute transport processes

Solutes may migrate in response to concentration gradients (diffusion), with flowing water (advection), or by a combination of these processes (Fetter 1993). The mass transfer that occurs by advection is a function of the concentration of solute, and the quantity of groundwater flowing. This is the dominant transport mechanism within aquifers, and sometimes within aquitards.

Contaminant transport may occur rapidly through fractures. Doser et al. (1998) found contaminants transported through a 3 m thick stiff clay within several days and Harrison et al. (1992) demonstrated that a significant contaminate plume could develop where groundwater velocities in the order of kilometers per year occurred through fractured clay. Aquifers that supply drinking water for 20 million people in Mexico City are vulnerable to contamination because of fractured lacustrine clay (Rudolph et al. 1991).

Diffusion

The dominant transport process in low permeability material is often diffusion, while hydrodynamic dispersion is generally ignored (eg. Love et al. 1995). Diffusion is defined as the movment of a solute in water from an area of greater concentration to an area where it is less concentrated (Fetter 1993).

Mass transfer of solute is proportional to the concentration gradient according to Fick's first law (Equation 3.6). The change in concentration with distance is described by Fick's second law (Equation 3.7).

$$f = -D\frac{\partial C}{\partial x} \tag{3.6}$$

Where F is flux (mols/m^2), D is diffusion coeffcient (m^2/s), C concentration (mol/m^3), and $\frac{\partial C}{\partial x}$ is concentration gradient.

$$\frac{\partial C}{\partial t} = D\frac{\partial^2 C}{\partial x^2} \tag{3.7}$$

Where $\frac{\partial C}{\partial t}$ is change in concentration with time.

The effective diffusion coefficient, D_e, is related to the diffusion coefficient for porous media, D_p and the free solution diffusion coefficient, D_o, by Equation 3.8 (Rowe et al. 1995).

$$D_e = \frac{D_p}{\phi_e} = \frac{D_o W_T}{\phi_e} \tag{3.8}$$

Where ϕ_e is effective porosity and W_T is the complex tortuosity factor which accounts for electrostatic interaction and adsorbed double-layer water, in addition to geometric factors.

Effective diffusion co-efficients in the order of 10^{-9} to 10^{-10} m^2/s have been used in studies of long term transport through clayey deposits (eg Desaulniers et al. 1982, Appelo & Postma 1996, Hendry & Wassenaar 2000, Konikow & Arevalo 1993). For the Geera clay, Hanor (1987) used a value of 10^{-9} m^2/s for the Geera clay and a tortuosity of 3 m/m.

Further detailed information on diffusive transport are available elsewhere (eg. Fetter 1993, Rowe et al. 1995, Appelo & Postma 1996). Discussion here is limited to case studies that highlight relevant diffusion issues.

The relative importance of diffusion relative to advection in solute transport within aquitards may be assessed by means of the dimensionless Peclet number, P_e (Equation 3.9).

$$P_e = \frac{VL}{D_e} \tag{3.9}$$

Where V is linear velocity, L is aquitard thickness, and D_e is effective diffusion coefficient.

For P_e<5, diffusion is the dominant process, while between P_e5–9 advection and diffusion is both important, and at P_e>9 advection alone is dominate. Diffusion dominates transport mechanisms where velocity is <0.001 m/year (Desaulniers et al. 1981). Model simulation of ^{18}O, ^{2}H and Cl contents in a Canadian till highlighted the importance of diffusion of low flow velocities. At velocities of <0.008 m/year diffusion is equally important to advective transport, and at velocities less than 0.001 m/year, diffusion dominates.

Fick's second law is solved for boundary conditions over geologic time for situations where geologic change caused chloride diffusion over hundreds of years. Diffusion of contaminants into clay at the aquifer-aquitard boundary may later re-enter the aquifer as concentration gradients reverse (Harrison et al. 1992; Rowe et al. 1995).

Diffusive transport may occur in a direction opposite to groundwater flow, as illustrated by the linear decrease of salinity downwards from the bed of Lake George where upwards seepage occurs Jacobson et al. (1991). Diffusive mixing was used to explain geochemical profiles in thick clay till in Saskatchewan, without envoking geochemical reactions (Hendry & Wassenaar 2000).

For fractured mediums dominated by advective flow, diffusion into peds may be a sigificant factor in retarding the transport of solutes. Examples of this process delaying breakthrough of contaminants in dual porosity rock aquifers and aggregated soils is provided by (Appelo & Postma 1996).

Table 3.6: Examples of aquitard effects on groundwater chemistry

Location	Nature of aquitard	Effects on underlying aquifer
Vale of York, UK [1]	glacial, lacustrine and fluviatile clays overyling sandstone aquifer	$>$ SO_4, $>$Mg
Atlantic coastal plain, New Jersey, USA [2]	wedged shapped, multi-layered, marine	$>$ SO_4, HCO_3 in confined aquifers
Vamsadhara valley, INDIA [3]	shallow alluvial aquifer overlain in some areas by upto 5 m of clay	$>$ NO_3 after monsoon in areas overlains by clay. Sorbed NH_4 is released as NO_3 when clay is saturated.
Mexico City, MEXICO [4]	lacustrine clay overlying aquifer	source of salts

1 Spears & Reeves 1975, 2 Pucci et al. 1992, 3 Rao 1998, 4 Ortega-Guerrero et al. 1997

3.3.3 Effects on adjacent aquifers

Clay aquitards can cause higher levels of TDS and other contaminants in underlying aquifers (Table 3.6). For example, glacial and lacustrine clays northwest of York, UK were found by Spears & Reeves (1975) to exert a significant effect on dissolved solutes of the underlying sandstone aquifer. Rao (1998) detected higher levels of nitrate below clay aquitards. This association was attributed to several mechanisms including the slower nitrification in clays versus sandy overburden, the volitalisation of nitrate in sandy material, the electrical membrane effects of clay and the absorption and later release of nitrate.

In some instances, aquitards provide protection to underlying aquifers from contamination and there are no measurable effects. For example, in the San Joaquin valley a regionally extensive lacustrine clay that is 6–36 m thick protects confined aquifers from induced leakage of shallow saline groundwater (Belitz & Phillips 1995). Salinisation of the confined aquifer has not been reported. Darcy flux through this layer was estimated to be 82–130 mm/year depending on groundwater pumping, which equated to a downward linear velocity about 300 mm/year. At that rate, it was estimated that saline groundwater would not arrive at confined aquifers for another 400–600 years.

3.4 Accounting for groundwater quality variability

The aim of groundwater quality monitoring in many instances is to detect changes that may be attributed to anthropogenic factors which may increase groundwater flux or contaminant loading. Valid interpretation regarding relative impacts requires a carefully designed monitoring program and interpretation protocol that accounts for all factors that contribute to variability (Keith et al. 1983).

Numerous references provide guidelines to assist design of appropriate groundwater quality monitoring programs (eg. Nelson & Ward 1981; Keith et al. 1983; Cantor et al. 1987), while others give instruction on appropriate statistical analysis (eg. Montgomery et al. 1987; Harris et al. 1987; Close 1989; Helsel & Hirsch 1992).

Equation 3.10 represents factors that contribute to observed groundwater quality changes (Loftis 1996).

$$y(t) = u(t) + \epsilon(t) + b(t) \tag{3.10}$$

For a number of sampling locations at time t, where y(t) is a vector of n groundwater quality observations, u(t) is a set of mean functions, ϵ(t) is a vector of noise, and b(t) is set of deterministic temporal trend functions.

Examples of noise terms include seasonal variation in a shallow aquifer and sampling error variance. The aim of groundwater quality monitoring is to identify and describe the trend function b(t), whether it be linear or step trend. A brief discussion of trend analysis as it relates to groundwater quality data is now presented.

3.4.1 Trend analysis

A definition of the term trend, in the context of water quality, is provided by Loftis (1996) .

> *"Trend is....a change in water quality over some period of time which is of interest from a management perspective."*

Identifying significant trends, and understanding causal processes is critical for appropriate management decisions. However, attention has historically focused on spatial 'snapshot' patterns so that are few data sets suitable for temporal trend analysis. This is a particular problem in regions with insufficient resources to support regular groundwater quality monitoring.

The b(t) function, from Equation 3.10 above, can be described either by:

- statistical significance testing eg. Mann-Whitney test, Kendalls Tau
- estimation of trend magnitude ie. studying cause and effect relationships

Though the former is more common and largely associated with regulatory approaches, the significance of a trend does not necessarily equate to the practical significance of a trend for resource management. An estimation of trend magnitude that describes cause and effect relationships is of far greater use to resource management.

Trend analysis for groundwater quality must address generally slow and persistent groundwater quality changes and the uncertainty of spatial characterisation (Loftis 1996). Slowly flowing groundwater means that serial correlation - where there is linear dependence between observations in time - is more pronounced. Samples need to be sufficiently well spaced in time to avoid this effect (Nelson & Ward 1981).

An appropriate sampling frequency is critical to testing significance because if a sample size is too large, statistical testing will be sensitive to changes that are too small for concern (Loftis 1996). Conversely, if the sample size is too small, statistical testing is insensitive to changes that require management action. It was found by Rajagopal (1986) for example, that a sample size of at least 50 was sufficient for characterisation of inorganic parameters such as chloride for Quaternary aquifers in Iowa. This was based on the estimation of the 95th percentile which was achieved within 6% accuracy by sample sizes of 50, 100 and 250.

Required monitoring frequency may be determined by deciding on an appropriate confidence level about the mean of the specified variable (Nelson & Ward 1981). The number of samples required where independence can be assumed (no serial correlation) is given by Equation 3.11 (Swan & Sandilands 1995).

$$n \geq (t_{\alpha/2,n-1}\sigma/P)^2 \tag{3.11}$$

Where σ^2 is population variance, n is number of samples, P is precision, α is confidence, and t is students distribution.

A useful trend analysis is difficult where the period of interest is longer than period of record (Loftis 1996). In this case, serial correlation is a more appropriate term than trend.

3.4.2 Well installation trauma

Well installation trauma refers to the variation in groundwater water quality after reactive substances are introduced or created in the subsurface by well construction. Drilling fluid and breakdown agents, and the placement of bentonite seals may potentially influence the chemistry of water, particularly in low permeability strata. The term 'well installation trauma' was coined by Walker (1983) to describe these effects, and popularised by Brobst & Muszka (1986), who reported the effects of three drilling fluids on groundwater chemistry.

Elevated chemical oxgyen demand (COD) persisted for 140 days for a well drilled with bentonite, and 320 days for a well drilled with guar gum. This was attributed to dissolved and suspended organic carbon, ferrous iron and ammonia in the breakdown additive, suspended minerals containing reduced elements (eg Fe, Mn, Cu) and natural organic matter. Elevated Cu, Pb and total organic carbon (TOC) was also noted.

Using additive breakdown components reduced COD demand for 50 days, but caused elevated sulphate. The end products of bacterially induced breakdown of the complex carbohydrate from guar gum would remain in the borehole vicinity unless removed by groundwater flow or well development.

The effects of bentonite seals in monitoring wells installed in low permeability strata was observed to persist for at least 6 years (Remenda & van der Kamp 1997). In this case, about half the concentration of the tracer Br, mixed with the sand-bentonite seal had leached to the underlying porewater after 6 years. Contamination of Cl, Na and SO_4 was also recorded. Bentonite well seals were noted by Brobst & Muszka (1986) to increase COD due to trace amounts of organic polymers and the long-term exposure of well seals to vigorous purging. Gypsum content of bentonite increased sulphate levels.

There are several means by which water samples contaminated by the drilling and installation process can be avoided: alternative drilling proceedures and improved piezometer design (Remenda & van der Kamp 1997; Wassenaar & Hendry 1999), elimination of suspect data, or correction of hydrochemical data by means of tracers (Kloppmann et al. 2001).

3.4.3 Groundwater vs. well water quality

The quality of water sampled in a pumping well whether it be an irrigation bore, domestic well or town water supply, is not always a reflection of groundwater quality (Schmidt 1977; Nightingale & Bianchi 1980). Differentiating between well water quality and groundwater quality is important where there are permeability differences, vertical stratification and lateral variability of groundwater quality and also where pumping rate has been variable or intermittent. Variability due to stratified plumes have been described by Cosler (1997).

Changes observed in wells of urban Fresno, California (Schmidt 1977) reflected the sources of groundwater within the well capture zone, particularly where flow occurs from contaminated sources including aquitards. The determination of long term quality changes must be prefaced by analysis of short term and seasonal changes to establish natural variability. Optimum sampling frequencies can then be achieved.

A subsequent study reported on the shapes of bivariate plots for EC versus pumping time and nitrate versus pumping time for urban wells in Fresno. Dramatic groundwater quality changes within the first 5 to 10 minutes of pumping were able to be grouped into 11 type concentration versus time curves. An exponential decrease or increase in EC or nitrate to a limiting or maximum value was the most commonly encountered pattern. The shape of the curves was related statistically to depth of screen intake, total well depth and the degree of contamination of upper zones of the aquifer system. A correlation was also found between EC changes and specific yield and the transmissivity of specific textural horizons, and the specific capacity of the well, which was itself dependant on drilling and construction methods.

The influence of near-field flow regime on the composition of groundwater near an irrigation bore was assessed by Ayers & Gosselin (1995) and Chen et al. (1998). The study was performed in a deep alluvial sand aquifer of the Platte River, Nebraska, approximately 20 m deep, bounded at the base by a 10 m thick silty clay. A total of 7 multi-level piezometers (total of 25 sampling points) equiped with bladder pumps were installed north and east of the irrigation bore at distances of between 1.5 and 11.5 m. Water samples were continuously collected over a period of hours and collection times corrected for lag times due to pumping.

In this field experiment, nitrate-N was found to be semi-constant where the flow regime was predominanty horizontal, and to change significantly where the depth interval sampled was in proximity to the top of the cone of depression as vertical flow components became important. Results were qualitatively analysed using a radial flow model that incorporated particle tracking (Chen et al. 1998). The area of influence (analogous to the steady-state capture zone) of the irrigation well during pumping was assessed. It was found that groundwater quality variation depended on the depth and length of the screen, and aquifer anisotropy which generally limited flow to horizontal pathlines.

3.4.4 Case studies of temporal changes

Selected examples of seasonal and long term studies of groundwater quality are now examined. A few studies have focused on establishing short term variability of groundwater quality and some of the medium and longer term studies include a seasonal analysis to establish background variability, as recommened by Schmidt (1977). Each of these studies will be briefly described, and key elements such as duration and sampling frequency is outlined in Table 3.7.

Seasonal changes

Seasonal changes are generally restricted to aquifers that are sufficiently shallow to respond to episodic recharge events or heterogeneity within the saturated and vadose zone. The relative dominance of spatial or temporal variability appeared to depend on site specific factors such as the degree of subsurface heterogeneity and the nature of contaminant sources. Many of these case studies illustrate the importance of heterogeneity which often dominates small scale investigations (Back & Baedecker 1989; Back et al. 1993). Groundwater hydrochemical heterogeneity has been demonstrated a centimeter scale (Ronen et al. 1987), over tens of meters (Bjerg & Christensen 1992; Jankowksi & Beck 2000), and at a regional scale (Spalding & Exner 1980).

Seasonal fluctuations in groundwater quality in the semi-arid Agra district of India were determined from 30 representative bores sampled pre-monsoon, post-monsoon and during the period of maximum pumping (Singh & Narain 1980). Seasonal fluctuations were determined to be of low magnitude, except for dilution of salt concentrations that occurred after the monsoon months of July to September. Variation in EC correlated with changes in Na and SO_4 and NO_3, and was greatest in high EC waters. No significant correlation was detected between any other major cations or anions. Analysis of samples from February, which coincided with the period of maximum groundwater pumping indicated increased concentrations of many major species relative to post monsoon sampling, for which no explanation was offered.

Groundwater quality variation in an Oklahoma aquifer were monitored on a weekly to fortnightly basis for 1 year by Hoyle (1989). The aquifer was a silty-loam soil beneath a 160 m^2 site within a residential area where nitrate fertilizer was applied to lawns. A total of 16 sampling points were installed in 4 clusters at depths of between 2.4 and 4.3 m. Significant spatial and temporal variability of hydrochemical parameters was observed. It was concluded that characterization of background water quality in this environment

Table 3.7: Selected case studies of groundwater quality changes over time

Location	Parameter	Trend	Land use[a]	Monitoring Years	Freq. [b]	No. wells
Seasonal studies						
Agra, INDIA [1]	TDS	Monsoon dilution	I	1	3x	26
Oklahoma, USA [2]	TDS, pH	Variable	D	1	W	16
Western DENMARK [3]	TDS, pH, NO_3	Small	D	1.3	M	80
Medium to long term studies						
West Virginia, USA [4]	Cl	Increase - decrease	D	4-60	IR	?
Fresno, California, USA [5]	TDS	Decrease	U			
Arkansas River Valley, Colorado, US [6]	EC	Stable	I	10	IR	30
San Joaquin Valley, California, USA [7]	Salnity, NO_3	Increase ?		I		
El Paso, Texas, USA [8]	Cl	Increase	D	15-45	IR	6
Tatiara, South AUST [9]	TDS	Increase	I	10	∼Q	6
Waiau, North Canterbury, NZ [10]	NO_3		I	9	M ?	5-13
Bow River, Alberta, CAN [11]	EC	Increase	I	20	M	20
Gigaree, N. Vict, AUST. [12]	EC	Variable	D	5	∼BM	31+13
Carson Desert, Nevada, USA [13]	TDS	Small increase	I	30	IR	3+?

[a] I=irrigated, D=dryland, F= forest, U=urban

[b] Q= quaterly BM = bimonthly M=monthly, W=weekly, IR = irregular

Sources: 1 Singh et al. 1980, 2 Holye 1989, 3 Bjerg et al. 1992, 4 Wilmoth 1972, 5 Nightingale & Bianchi 1977, 6 Konikow et al. 1985, 7 Schmidt & Sherman 1987, 8 Hibbs 1998, 9 Stadter et al. 1989, 10 Close 1987, 1989, 11 Beke et al. 1993, 12 SKM 1995, 13 Maurer et al. 1995

Table 3.8: Magnitude of groundwater salinity increase in selected areas

Location	Period	Δ Salinity (mg/L)	Rate (mg/L/year)	Comments
Sevier Desert, Utah, USA [1]	1959- 1969	550- 950	+40	irrigation water recycling
Pavant Valley, Utah, USA[1]	1958- 1984	1700- 6500	+185	lateral inflow
Beryl-Enterprise, Utah, USA [1]	1960- 1983	∼200- ∼1100	+40	irrigation water recyling
Tamil Nadu, INDIA [2]	1985- 1991	∼125- ∼830*	+710*	lateral inflow due to pumping
El Paso, Texas, USA [3]	1980- 1995	∼50- ∼300*	+20*	drawdown of saline water near river, lateral inflow and possibly upconing
Tatiara, South AUST [4]	1978- 1988	∼3500- ∼4900	+134	predominately irrigation recycling
"	1982- 1988	∼7200- ∼8200	+321	worst case - largest increase in most saline gw
Carson Desert, Nevada, USA [5]	1960s- 1992	90- 110*	+0.7*	possible upconing

* as Cl

Sources: 1 Moody et al. 1986, 2 Sukhija et al. 1998, 3 Hibbs 1998, 4 Stadter & Love 1989, 5 Maurer et al. 2000

required several strategically located nested piezometers, and non-parametric statistical analysis due to non-normal distribution of parameters.

Large temporal and spatial variation in groundwater quality was observed in a relatively homogeneous shallow sandy aquifer in Western Denmark (Bjerg & Christensen 1992). Major ions, pH, alkalinity and nitrate was monitored over 15 months in 80 wells at a small experimental site in agricultural land. Depth specific groundwater monitoring was required to identify significant variations with depth due to varying quality of agricultural recharge waters. Observed changes with depth were much more important than temporal variability over a period of months. Horizontal well spacing of 10 m was required to fully characterize horizontal groundwater quality variation in this homogeneous environment.

Medium to long term trends

Monitoring and evaluation of groundwater quality trends over the medium to long term (years to decades) is a relatively recent science that partly reflects increasing awareness of the importance of sub-surface water resources. Examples of studies that detected increasing salinity over time are listed in Table 3.8.

A classic study by Close (1987,1989) identified the effects of irrigation development over the medium term, where 9 years of monthly data meant that it was possible to deseasonalise data prior to trend analysis. However, many studies have been hampered by technical deficiences such as irregular monitoring data.

Some long term studies from North America were reported in a special issue of the Journal of Irrigation and Drainage Engineering (Bouwer 1987). In the Bow River Irrigation district groundwater EC increased at 6 of 9 sites, with a marked increase in scatter over time indicating more variability due to irrigation practices (Beke et al. 1993).

Annual increases in groundwater salinity up to 320 mg/L per year were reported from the Tatiara irrigaition area in South Australia by Stadter & Love (1989). Greatest rates of groundwater salinity increases were associated with higher salinity waters, though salinity decreased in some wells.

The Girgarre salinity control project aims to lower the watertable in a salinized part of the Shepparton Irrigation area (SKM 1995). It consists of 3 multiple well point abstraction sites, and a disposal basin. Groundwater quality in the shallow Shepparton Formation was monitored over 6 years with salinities either increasing or stabilising. Pumped groundwater salinity has increased from 2 000 μS/cm to 5 000 μS/cm at one site, and 5 000 to 9 000 μS/cm in a second bore.

3.5 Causes of groundwater quality variability

A cause-effect relationship between two factors such as groundwater level and quantity, is not proved simply by an association or correlation (Iman & Conover 1983). Four conditions must be satisfied in order to prove a cause-effect relationship:

- correlation - at a significant level in terms of management or statistical criteria
- consistency - occuring at >1 site or bore, or with each similar event in time
- responsiveness - may depend on time lags
- mechanism - physical process

Environmental assessments, much less groundwater studies, will not often achieve all of these criteria due to limited spatial and temporal scope. Nevertheless, there are a number of physical processes of both natural and anthropogenic origin which clearly do impact on groundwater quality. These will now be discussed with reference to the case studies outlined above.

3.5.1 Episodic recharge

Understanding background fluctuations in groundwater quality is a prerequisite to quantifying effects due to anthropogenic causes. Recharge, which occurs episodically in many areas of the world, may significantly alter groundwater composition where aquifers are shallow and permeable flowpaths or macropore bypass channels occur.

Cyclic changes, as defined by Pettyjohn (1976, 1982) are rapid and significant changes in groundwater quality in the absence of anthropogenic factors. Detection of cyclic changes in groundwater quality requires vertically discrete sampling and sampling on a daily to weekly frequency. The definition of cyclic groundwater quality arose from groundwater investigations beneath Ohio oil-field brine ponds, where contaminants were intermittently re-introduced to the groundwater system despite the cessation of disposal. The danger of assuming constant 'background' contaminate levels was obvious. For cyclic groundwater quality fluctuation to occur, unsaturated zone storage and macropore flow conditions and seasonal recharge events were necessary.

Episodic flushing of salts stored in the vadose zone in arid zones was studied by Drever & Smith (1978). Repeated cycles of evaporation, followed by resolution was found to significantly change hydrochemical composition, due to differences in the rates of dissolution. Rapid dissolution of Na and Cl contrasted with slower rates for carbonate and SiO_2, leading to solutions depleted in SO_4, Mg and SiO_2 relative to Cl. Such a phenomena was noted to potentially affect hydrochemical mass balance calculations.

Recharge events may be recognised by examining the frequency distribution of temporal EC data (Hoyle 1989). In that case, a trimodal distribution of EC data was evident in a shallow soil aquifer due to discrete rainfall events.

3.5.2 Agricultural development

Variable croping and vegetation

Landuse changes may impact groundwater quality of shallow aquifers for a variety of reasons. These include changes to recharge regime, fluxes of fertilisers and pesticides associated with specific crops, and also hydrochemical effects within the root zone. These changes may be observed over time, or more often, as comparisons between adjacent paddocks under different landuse.

For example, Hoyle (1989) observed elevated EC, CO_2 and HCO_3 levels close to vegetation that was attributed to respiration and associated microbe activity. The hydrochemical effects of this activity were very localised however, since these effects were not detected a few 10's of meters away.

Beke et al. (1993) reported that groundwater below perennial crops was characterised by a higher long-term mean EC, than that from sites with annual crops.

The magnitude of groundwater salinity changes over a 6 year period near a saline disposal basin were found to be dependent on landuse (SKM 1995). The greatest salinity decrease occurred in piezometers located below pasture, compared with small changes beneath dryland and annual crops.

Flushing and dilution by deep drainage

In some instances, irrigation practices have led to an improvement of groundwater quality due to dilution by increased deep drainage. Increased recharge, or deep drainage, may occur either under dryland or irrigated cropping. One of the most difficult problems in hydrogeology is to predict movement of a contaminant in ground water when concentrations are decreased by dilution (Wilmoth 1972).

Dilution of groundwater by irrigation return flow may result in less saline groundwater, a factor which is more pronounced under annual compared with perennial cropping (Beke et al. 1993). The subsequent establishment of a vertical gradient in groundwater quality was dependent on the type of aquifer material. Seasonal dilution effects were noted in the Bow River irrigation area by Beke et al. (1993), an area which recieves irrigation water from the river. Depth affected by seasonal deep drainage was estimated at 1.5 m.

Close (1987) monitored groundwater quality of a shallow unconfined aquifer before and after implementation of a irrigation scheme, observing an increase in depth of drainage of irrigation water over time. A drainage depth of approximately 9 m was determined from monthly vertical profiles of major ions and nitrate to a depth of 30 m acquired over a 3 year period. Nitrate flux was found to increase with irrigation, but concentrations decrease due to dilution. In the absence of fertilizer application, nitrate fluxes were attributed to animal wastes. Decreasing NO_3 concentrations were attributed to increased irrigation drainage that occurred in Waiau Irrigation Scheme, north of Canterbury, South New Zeland (Close 1987). This trend was established on the basis of 8 years of monthly data that were rigorously analysed by statistical techniques. The depth of drainage was determined based on the average depth of upper and lower concentrations.

Decreasing chloride concentration over time under irrigated land in Portugal could be attributed to flushing of contaminated groundwater (Stigter et al. 1998). Flushing of groundwater may be achieved artificially by implementing enhanced recharge schemes. One such project in Fresno, California successfully lowered groundwater salinity at distances up to 1.6 km from the infiltration basin (Nightingale & Bianchi 1977). This innovative project was able to secure resource access for the domestic and industrial users who rely soley on groundwater, by recharging the aquifer with relatively fresh river water.

Re-cycling of irrigation water

Under some circumstances, deep drainage from irrigation contributes additional salt to the groundwater system. This occurs if irrigation waters are partially saline, and the water use efficiency (WUE) is high and conversely, that the leaching fraction is low. Consequently, increased salt load may not be balanced by increased fluxes.

For example, Stadter & Love (1989) provide an example of lucerne irrigated with 3 000 mg/L water. A WUE of 50% and 70%, would result in drainage salinity of 5 500 and 8 000 mg/L respectively. If groundwater, rather than surface water were used to irrigate, the cycle may be repeated, causing increased salinity over time.

Downward leaching of salts remaining after irrigation, was implicated as the primary cause of groundwater salinity increases in the Tatiara irrigation area (Stadter & Love 1989). Some salt may also have been sourced from upconing, lateral migration and preferrential flow through leaky bore installations, but increases occurred without major changes in groundwater storage, or groundwater level decline.

A number of other irrigation areas around the world have suffered similar effects. A small, but longterm increase in salinity of groundwater just below the watertable in the Bow River Irrigation area, Alberta was attributed by Beke et al. (1993) to improvements in irrigation efficiency. Deeper groundwater was apparently not affected in this case. In the San Joaquin Valley, shallow water table areas were found to be particularly sensitive to the effects of irrigation return flow.

3.5.3 Groundwater flow enhanced by pumping

Groundwater pumping causes increased groundwater fluxes. The importance of lateral versus vertical (either upwards or downwards) fluxes depends on the relative permeability of each flowpath.

Increased lateral flow

Maurer et al. (2000) reviewed available literature and several USGS reports on the effects of irrigation on groundwater quality in the Carson Desert Agricultural area of Nevada. The irrigation area began in 1903 in a basin with three sedimentary aquifers of varying groundwater quality. Near-surface clay inhibits vertical flow in parts of the basin.

A comparison was made between irrigated and non-irrigated areas and lateral groundwater flow versus discharge areas to determine whether irrigation affected groundwater quality. The potential affects of reduced irrigation was also assessed. The conclusion was that irrigation had only a minor effect on shallow groundwater quality to a depth of approximately 20 m. Chloride concentrations were found to increase from 90 to 110 mg/l over 30 years, a range greater than analytical uncertainty and attributed to decrease in hydraulic head causing lateral flow of saline water from adjacent aquifers.

Increased vertical flow

Upconing of saline or brackish groundwater may occur in stressed aquifer systems. There are numerous examples of this process occuring in coastal environments, but more rarely in continental settings. For example, Figure 3.1 depicts saline intrusion on the Atlantic Coast of the US via preferrential flow paths (USGS 2001). The only example of downwards flux comes from Australia (Stadter & Love 1989).

The subsurface salt distribution is one factor that determines whether increased vertical flux degrades groundwater salinity. In some environments, groundwater quality improves with increasing depth (eg. Spalding & Exner 1980). Typically however, groundwater quality deteriorates with depth (eg. Close 1987; Beke et al. 1993), particularly in coastal areas.

Increased chloride levels coincided with increased pumping in the El Paso area of Texas (Hibbs 1998). Salinisation was attributed to the thickness of saturated sediments, hydrochemistry of the saturated basin fill, distribution and continuity of mud interbeds and density of saline water. Modeling studies of the area concluded that horizontal to vertical anisotropy in the basin as a result of mud interbeds could prevent salinization due to upconing of brackish waters.

Upconing of saline groundwater is occuring due to pumping of the Great Prairie Bend alluvial aquifer in Kansas (Sophocleous & Ma 1998). Management is difficult, in part because of uncertainty regarding the distribution of low-permeability clay lenses. A groundwater decision support system was developed that considered the size of clay layer discontinuities as one of 15 variables on which the probability of salinisation at bore sites was estimated.

3.5.4 Water quality improvement after pumping ceases

If groundwater pumping were to cease, groundwater salinity may decrease due to influx of fresher water. This is an unlikely scenario in agricultural areas, but an example is

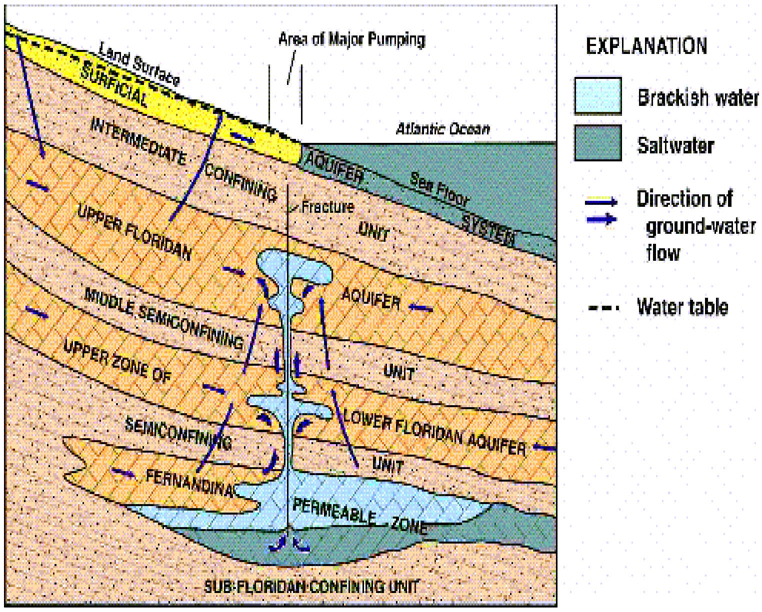


Figure 3.1: Saline intrusion along preferential flowpaths in a multi-layered aquifer, Atlantic coast, US (Source: USGS, 2001 after Spechler 1994)

reported by Wilmoth (1972) after industrial pumping in an West Virginian oil-field was discontinued. In one well, chloride concentration decreased over a decade, a much shorter period than the increase which had occured over four decades. At another site, chloride increased from < 100 mg/L to $\sim$3 000 mg/L in 2 months, and then 2 months after pumping ceased had dropped to 1 000 mg/L. Chloride levels continued to decrease over the next few months, and were only slightly higher than pre-pumping values 2 years later. At another site, chloride levels increased from 55 to 1 900 mg/L occurred over a period of 5 years, yet took 10 years to return to natural levels after pumping ceased, indicating that groundwater quality effects can be persistent.

Bore casing leakage

Abandoned test holes and leaky boreholes can result in significant cross-contamination of aquifers. During the several decades that have elapsed since aquifer were developed as water supplies, the life expectancy of many bores (30–50 years) has expired, resulting in an estimated 30 000 leaky boreholes in local area (Lacombe et al. 1995). In West Virginia, vertical movement of salt occurred extremely slowly in undeveloped areas, but in other areas rapid vertical leakage occurred predominately via unplugged wells and test bores Wilmoth (1972).

Ramamurthy & Holmes (1983) reported on investigations in the Angas-Bremer irrigation area of South Australia where trends of increasing salinity were noted in confined aquifers underlying water table aquifers with salinities upto 20 000 mg/L. The presumption that induced vertical seepage from the unconfined aquifer caused this trend was complicated by enigmatic decreasing trends. Such trends were noted in adjacent boreholes and the possibility of lateral flow of saline groundwater was raised. Furthermore it was noted that the spatial distribution of salinity in the confined aquifer was a poor indicator of possible vertical leakage, since the unconfined aquifer was not always more saline. The study concluded that regional scale vertical leakage was unlikely, though point source seepage through faulty, corroded and leaky borehole casings was likely.

In the San Joaquin valley, it is estimated that bulk K has increased, despite reductions in matrix K due to compaction of clayey aquitards. Increased K_v was attributed to over 100 000 wells constructed with long screens and perforated casings that enable mixing of groundwater from different layers (Moody et al. 1986). Hibbs (1998) cites an example of nitrate contamination that occurred when a shallow pool of urban runoff containing fertilizers 'cascaded' down an abandoned well into the deep aquifer.

Based on numerical modelling, Lacombe et al. (1995) found that contaminated water can rapidly migrate down leaky boreholes even if the bore is filled with aquifer sediment. An entire contaminant plume, or significant portion of it could be diverted to a lower aquifer via an open borehole at high (1.0) gradients. Total flow rate through the borehole was very sensitive to borehole radius ($Q \alpha r^4$).

3.6 Towards an integrated approach

Advances in understanding of complex sub-surface systems, and the potential impacts of agricultural activities including groundwater abstraction require a multi-disciplined approach. Groundwater quality variability nor water-matrix interactions should be considered in isolation. It is rare that investigations bring together techniques and insights from hydrology, chemistry, geotechnical engineering, geology and geophysics and mathematics (in the form of numerical modelling). Yet, such a conjunctive approach is more likely to converge on a realistic findings, particularly where individual strands of investigation proceed in a iterative fashion.

3.6.1 Testing flow models with hydrochemical information

Integration of both hydrogeochemical and hydrodynamic data is an emerging field, as demonstrated by a recent workshop on the use of hydrogeochemical information in testing groundwater flow models (Bath & Lalieux 1997). The general concensus of this workshop was that hydrogeochemical data could improve the credibility of site-specific groundwater flow models. However, the complexity of processes in 3D and the number of unknown parameters limiting testing of flow models with hydrochemical information to semi-qualitative measures. Qualitative direct linkage was best suited to near-field simple cases of 1D or 2D transport where relatively rapid reactions occur in a stressed environment.

In low permeability environments, mixing of groundwater masses was found to be reflected in hydrochemistry more sensitively than hydrodynamic measurements. Furthermore, some hydrogeochemical parameters are records of cumulative changes while others are indicative only of local water-rock interactions over short time scales. But hydrochemical data have proved useful in identifying boundary conditions, and rapid flow pathways and hydraulic information can be checked with isotopic transit times.

Complex groundwater management problems may be addressed by combining groundwater hydraulic models, solute transport models and optimisation techniques (Gorelick 1983). While groundwater hydraulic models are now a common tool, the joint use of groundwater simulation models and optimisation methods have not been widely applied, and in the Murray-Darling Basin, have been developed only in the Namoi catchment (N. Merrick, pers. com.). However, hydraulic models are currently limited by parameter uncertainty (Gorelick 1983), a shortcoming which is particularly relevant where contaminated recharge or migration of saline groundwater is a concern.

Linkage of optimised hydraulic models with groundwater quality models is even less common. Gorelick (1983) documents a number of case studies which illustrate different numerical methods of linking such models, but point out that groundwater quality management models are limited to linear problems where the groundwater flow field is well defined. Methods to solve non-linear problems, such as multicomponent chemical interactions in solute transport and saltwater intrusion problems where density differences are a significant driving force are yet to be developed.

3.6.2 Groundwater pressure vs. quality equilbrium.

Steady-state groundwater pressures do not necessarily correlate with constant hydrochemical parameters. A number of studies have shown that groundwater quality degradation may occur prior to, or independently of significant groundwater level changes that indicate pressure disequilibrium. This is particularly the case if the cause of degradation is related to variable contaminate loading rather groundwater flux (eg. irrigation re-cycling).

The dangers of predicting long term salinity changes based on short term data is highlighted by Konikow & Person (1985) who recalibrated a model which had earlier predicted a 2–3% increase in groundwater salinity, when no significant increase in salinity was actually detected over a 10 year period. It was concluded that the original model could have correctly predicted equilibrium even within the range of earlier parameters, though no sensitivity analysis was conducted at the time. The year long calibration data had coincided with a perturbation in a long term dynamic equilibrium. Groundwater salinity had reached an equilibrium with irrigation within a period of 12 years, or flushing by 3 pore volumes of groundwater. This also implies that while groundwater quantity data may indicate equilibrium, groundwater quality may take longer to equilibrate.

In the Tatiara irrigation area , increasing salinity was not associated with changes in groundwater storage, so from a quantity basis it would appear that groundwater pumping was sustainable (Stadter & Love 1989). Management of groundwater and improvements in irrigation efficiency would therefore at best, only limit groundwater salinity increase in such a susceptible environment.

3.7 Summary

- Aquitards are important components of the groundwater system which can exert a significant influence on the hydrology and chemistry of aquifers. Scale dependent permeability determines important vertical flux mechanisms, whether by diffusion, matrix flow, or by rapid flow pathways such as aquitard windows and leaky bore casings.

- Groundwater quality varies over the short and long term in response to many natural and development related factors. A cause-effect relationship may be established by identifying a correlation where a response occurs consistently in similar situations. For groundwater quality variability, the physical mechanisms that control solute transport must also be determined.

- Due to logistical challenges, temporal changes in groundwater quality are often ill-defined and limited to localised, short term examples. Groundwater quality has improved with agricultural development in some situations, but in others, groundwater quality degradation may occur prior to, or independently of groundwater level disequilibrium, due to changes in contaminant flux.

Chapter 4

FIELD AND LABORATORY TECHNIQUES

In this chapter the methods used to acquire new hydrochemical and hydrogeological information are described. Several weeks of fieldwork, months of laboratory testing and years of monitoring were a precursor for developing realistic hydrodynamic and hydrogeochemical models.

4.1 Hydrogeology

4.1.1 Geophysical investigation

Borehole conductivity and gamma logging

Gamma ray activity and bulk electrical conductivity were obtained using a GEONICS EM-39 borehole logging device (McNeill 1986) (Figure 4.2B). First, gamma ray activity was run at a speed of 0.05 m/s using a time constant of 8 seconds. Then, an electromagnetic induction logging tool was then run at a speed of 0.50 m/s to produce a bulk electrical conductivity (ECa) log. Data was sampled at 25 mm depth increments.

Electrical imaging

Electrical resistivity images were obtained at the Rice Mill and Tubbo sites using a galvanic technique (Acworth 1999).

Two multi-core cables were used at each location with the resistance measuring instrument (ABEM SAS 4000) located at the mid point between the cables. Each cable had 25 connectors at an electrode separation of 5 m, with a total image line length of 245 m. The configeration of electrodes gave detailed coverage of the apparent resistivity image between depths of 2.5–15 m and less intense coverage between 20 and 40 m.

At Tubbo site, image lines were located adjacent to the bore site (Figure 1.3), trending in a north-south direction (EI-1, Aug-99), and parallel to this, 200 m towards the DLWC observation bores and offset 50 m toward the north (EI-2, Feb-01). A third image line (EI-3, Apr-01) was located parallel to these, but located 1.8 km away, and centered on the DLWC observation bores. The first image was obtained using a manually switched patch box, and the remainder using the ABEM LUND automatic switching device.

An inversion routine was used to derive a forward model image that matched the field data image (Acworth 1999). The EC values and depths of the model distribution are considered a solution to the field data.

4.1.2 Drilling, coring and piezometer installation

Drilling was carried out in June, 1998 at the Tubbo and Gundaline sites, using a top-drive table rotary mud drill rig was operated by Tony Sander of GeoEng drilling. A 160 mm diameter bladebit was attached to HW (115 mm OD) drill rod, with HQ (98 mm OD) casing and a triple barrel mechanism as described below.

Three holes were drilled at Tubbo, 35 m to the east of the irrigation bore GW59113. Minimally disturbed core were recovered from the first hole, which was completed as a multi-level groundwater monitoring piezometer with a total depth of 96 m. Effort was directed towards improved piezometer design, within time and budgetry constraints. A maximum of two, rather than three piezometers were installed per hole and the depths of screens within each hole carefully planned so that the vertical spacing, and hence the number of bentonite seals between each sampling point was maximised. Piezometers were installed using class 18 (65 mm OD) PVC pipe with screw fittings and O-ring seals (see construction details in Appendix).

Core recovery

A triple barrel coring system, developed at WRL (Figure 4.1) was used to obtain a minimally disturbed 100 mm diameter core in 1 m lengths through the clay horizons. This was achieved with a core barrel extending in front of the drill bit, with an inner barrel fitted inside and protruded through the bottom cutting face of the drill bit by 30 mm for clay and 5 mm for sands. Contact between the drilling fluid and the surface of the sample was limited to fluid displaced upwards through a footvalve. A latch mechanism located the sample barrel inside the drill pipe and enabled the sample barrel to remain free from the rotation of the drill bit and allowed for easy recovery of the sample barrel via the use of standard wire line recovery techniques.

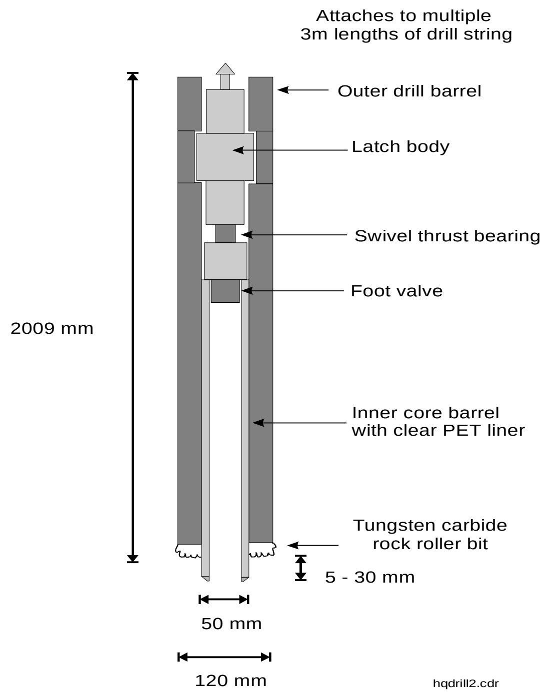


Figure 4.1: HQ triple barrel system for recovery of minimally disturbed cores.

A core catching device constructed of stainless steel was installed inside the core barrel at the base of the clear PET (polyethyleneterephthalate) liner, which enabled visual inspection of cores on site. The cores were extruded from the core barrel while inside the PET liner and this prevented loss of pore fluid and significant chemical alteration of the core. The cores were capped top and bottom, sealed and kept in cold storage.

Approximately 44 m of core were obtained from each of 3 clay sections (0–32, 45–64 and 75–83 m), which provided the source material for the detailed laboratory analysis.

Purging and development process

Mud additives were required to drill at depth, despite the potential for groundwater contamination. The drilling fluid used at Tubbo was Bariod Biopolymer^{TM}, an organic colloid containing trace amounts of paraformaldehyde and sodium hydroxide.

Recovery of drilling mud was enhanced by lowering the viscosity of drilling muds to near water with a dispersing agent Breakback^{TM}. Removal of residual drilling polymer with aided by purging with Bariod Barachlor (NaOCl). Barachlor was mixed with water a concentration of 100 mls per 200 L, and siphoned down each peizometer, allowed to diffuse and react with drilling fluid for a week and then removed along with excess groundwater

Table 4.1: Details of pressure transducers deployed at Tubbo site on the 7-Aug-98.

Bore	Transducer SN	Transducer rating (bar)	Transducer sensitivity	Cable length (m)	Depth (m)	Water level range (m bc) [a]
Tub B1	1027425	1	10.021	30		
Tub B2	1045393	1	10.03	30	26	14.8 - 25.7
Tub B3	1045426	1	10.001	30	26	18.2 - 22.9
Tub B4	1060177	1	10.022	30	26	15.2 - 26.1
Tub B5	1088419	2	4.974	30		17.8 - 24.6
Tub B6	1088424	2	4.981	40	36	17.8 - 25.8
30282	1045420	1	10.059	30		
30284	1065403	1	10.056	30	∼27	15.8 - 26.7

[a] mbc=meter below casing

with a GRUNDFOS MP1 pump.

The piezometers were developed immediately following the purging process, several weeks later by airlift pumping (courtesy of DLWC), and again in August when field parameters were monitored. Hydrochemical sampling was delayed as long as possible prior to the beginning of the irrigation season in September.

4.1.3 Pump testing and permeability analysis

Groundwater level data collection

CTL Hydrokit loggers equipped with DRUCK pressure transducers, were installed in each of the new piezometers and two existing DLWC monitoring bores, 1.8 km away (Table 4.1). Calibration of these loggers was checked prior to deployment, and where necessary re-calibrated using a DRUCK pressure gauge and Equation 4.1 with the assistance of M. Howden (DPW Manly Hydraulic Laboratory).

$$Sensitivity = full\,range\,output\,(mV)/10/\,transducer\,sensitivity\,rating \qquad (4.1)$$

The loggers were configured and primed to collect data in multi-phase pump test mode (AS 1990) from the start of the pump test, 8:30 am, 11th September, 1998. The CAT motor which drove the turbine pump was started and warmed up, and at 8:30 am, the turbine pump was activated. After the first pumping cycle, CTL loggers were programmed to record data at 6 hourly intervals (6 am, 12pm, 6 pm, 12am).


A


B


C


D

Figure 4.2: Groundwater field methods. **A** Piezometer and HYDROKIT water-level logger setup at Tubbo site. **B** Downhole geophysical EM-39 logging. **C** Pumping piezometer with GRUNDFOS pump, through flow cell for measurement of unstable parameters. **D** UNSW Groundwater Centre, mobile chemical laboratory.

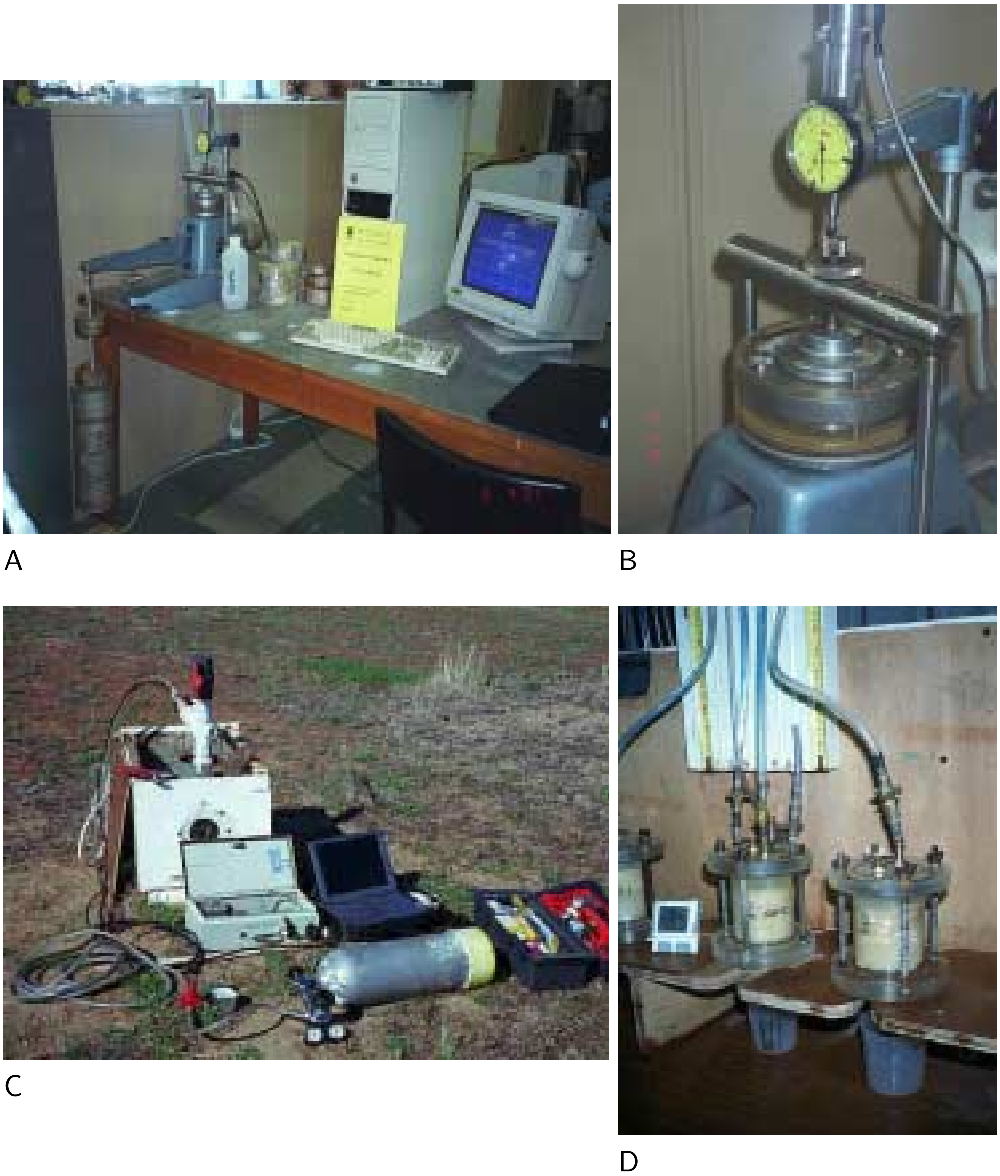


Figure 4.3: Permeability testing methods **A** Oedometer testing of consolidation. **B** Oedometer testing of core sample 2 cm thick by 7 cm diameter. **C** Slug testing of piezometers using compressed air. **D** Permeability testing of core samples by falling head permeater.

Slug testing of piezometers

In-situ horizontal hydraulic conductivity was determined by a piezometer slug test (Figure 4.3C). Head recovery in a piezometer was automatically measured after a near instantaneous change in head caused by the release of compressed air. The piezometers were pressurised to between 10–40 kPa, and the water level allowed to equilibrate prior to releasing the shut-in valve at the surface. Each test was repeated several times at varying pressures.

More detailed information regarding theory, execution and interpretation of slug testing is provided by Butler (1998). It is noted here that because a very small volume of the formation is tested, the method is sensitive to the effects of drilling such as poor development which may result in a low permeability skin surrounding the screen. The modified slug test for low permeability formations (Neuzil 1982; Bredehoeft & Papadopulos 1980) was not required for piezometer B5 in the Calivil aquitard due to rapid hydraulic head recovery. The standard Hvorslev method (Kruseman & de Ridder 1990) of interpretation was adopted.

Falling head permeameter testing

Falling head permeameter method is described by Holtz & Kovacs (1981) as reliable for hydraulic conductivity in the range 10^{-7} to 10^{-11} m/s. Permeameter apparatus and methodology were modified from that described by Crawford (1997) (Figure 4.3D). The falling head test board consisted of 8 mm ID clear plastic tubing constrained by wooden guides to prevent changes in tube volume. The test cells were made from clear perspex screwed into top and base plates that were separated by rubber 0 rings. A valve on the top allowed expunging of excess air from the chamber prior to testing. The base plate consisted of a plastic filter that was 50% open area and a fine mesh to prevent loss of core material. Fine gravel or sand separated this base plate from the sample. The outlet was regulated by a flow valve.

Minimally disturbed core samples of radius 0.055 m (or trimmed to 0.035 m for narrow cells) and 0.1–0.12 m high were mounted using bees wax in clear perspex cylinders. These were then mounted in the chambers, the seal of each cell checked, and the sample inundated in a water bath until hydraulic equilibrium was achieved. This process typically lasted 1 week to 1 month for clayey silt and half an hour for clayey sand.

Deionized water was used as the permeant for test runs, followed by groundwater that was obtained from piezometers at the Tubbo site (shallow aquifer, B1, for clayey sand

and middle aquifer, B4, for Lower silt unit). The test water was boiled and allowed to cool so as to remove dissolved gases which are known to reduce measured permeability (Collis-George & Yates 1987). Evaporation from the top of the falling head apparatus was considered to be insignificant as testing was completed within minutes to hours.

A hydraulic head of 1–2 m (equivalent to a gradient of between 10 and 20), was established across the sample tube. Water temperature was monitored concurrently with head measurements and corrections for viscosity variation due to changing temperature were applied according to Lambe (1951). The effects of 1-2 mm of sample swelling or consolidation during testing were considered to be negligible.

Oedometer testing

Geotechnical testing of core samples was undertaken with a standard fixed ring Casangrade cantilever style oedometer (or consolidometer), according to the method of Bowles (1978) at the UNSW School of Civil and Environmental Engineering Geotechnical Laboratory (Figure 4.3A,B). Briefly, the coefficient of consolidation c_v, was calculated based on the time for half the total settlement to occur at each load increment. The coefficient of compressibility a_v, was determined based on moisture content and void ratio, was calculated based on wet sample weight before and after settlement, and final dry weight.

Consolidation tests were performed on minimally disturbed core which were carefully mounted within a 75 mm diameter, 20 mm high fixed ring. The sample and ring were then installed in a fixed-ring consolidometer, with porous stone top and bottom to permit drainage and inundated with deionized or groundwater from the study site. Incremental loads, of an order expected under field conditions were applied, beginning with a 24 kPa pre-settlement load. Subsequent loads of 400, 800 and 1 600 kPa were applied. A digital gauge and PC computer were used to automatically record changes in thickness of the sample due to settlement from 0.1 to 10 000 seconds. This phase of testing typically occurred over a period of one week for each sample.

4.1.4 Solid phase analysis

Methods employed for mineralogical analysis and determination of basic physical properties (eg. particle size, porosity) of the clayey aquitards are described by Timms & Acworth (2002b) which is available on the accompanying CD-ROM.

Thin section examination of indurated clayey sand.

Thin sections of the indurated clayey sand, mounted on microscope slides, were prepared by R. Flossman (UNSW Geology) using a modified version of the method described by Flossman (1999). Specifically, a solvent dye, called 'Waxoline Blue APFW', produced by 'Kraftcolour' was mixed with Araldite D Resin in the ratio of 1% by weight. The core sub-samples were impregnated with this mixture in order to highlight pore structure. Samples were then prepared as 0.03 mm thick thin-sections mounted on microscope slides by standard methods.

The thin sections were examined under a Leitz polarising light microscope (Lewis & McConchie 1994). Magnification powers of up to 50, 125 and 500 times were used with fields of view of 4, 1.6 and 0.4 mm diameter respectively. The diameter of porethroats was estimated with the aid of an eyepiece graticule. Photographs were taken by attaching a modified SLR camera, capable of remote firing, to the microscope.

Dominant mineralogy and cement type were identified using standard petroscopic techniques outlined by Gribble & Hall (1985) with specific optical properties (outlined in supplementary tables included on CD-ROM). Although Younger (1992) used 300 point counts, the number of points counted to obtain consistent results for these samples was found by trial to be greater than 500. Mineral identifications for each slide were subsequently determined at 1200 'count' points.

An estimation of the effective porosity was made based on average diameters of minimum pore throats openings using the method of Younger (1992).

4.2 Hydrochemistry

4.2.1 Sampling procedures

Analysis of unstable parameters

As the standing water levels were at approximately 20 m depth, a submersible pump (GRUNDFOS MP1) was required to recover water samples. Discharge from the pump was fed through a flow cell equipped with ORION electrical conductivity (EC), pH, dissolved oxygen (DO), redox and temperature probes (Figure 4.2C). This proceedure minimised degassing associated with change in pressure, and ensured a constant flow (>30 cm/minute) of water past the DO electrode to prevent localised oxygen depletion.

Electrodes were calibrated according the proceedures outlined in Table 4.2 and reported according to the following conventions with corrections as required. EC mea-

Table 4.2: Field electrode calibrations

Electrode	Calibration
pH	twice daily, 2 point calibration with pH 4, 7 buffer solutions
DO	deionized H_2O, aerated by shaking with air, periodic zero check with saturated solution of sodium metabisulfite
Eh	twice daily, Zobell's solution with Eh +232 mV
EC	temperature corrected check with at least 2 standard EC solutions

surements were temperature corrected to 25^0C. A salinity correction was not applied to measured DO given the relatively low salinity groundwater encountered. Eh measurements from the Ag/AgCl electrode (E_0) were converted to the standard normal hydrogen electrode (E_{NHE}) using the conversion factor in Equation 4.2 and the fact that a 1 mV increase occurs for each 1^0C decrease in temperature.

$$E_{NHE} = E_0 + 204@20^0C \tag{4.2}$$

Water samples were taken for analysis when these parameters stabilised, to ensure that stagnant casing water was removed, and that representative groundwater samples were obtained.

Bicarbonate, HCO_3, was determined by titration of an aliquot of sample with 0.01 M HCl down to an endpoint of pH 4.3. Acidity due to free carbon dioxide, CO_2(aq), and undissolved H_2CO_3, was measured by titration of a sample aliquot with 0.025 M NaOH to pH 8.3. For waters with pH > 8.3, carbonate, CO_3^{2-} was determined by titration with HCl down to pH 8.3. These titrations were performed with an ORION meter that was fitted with a Ross Sureflow pH electrode to ensure rapid equilibration. The occasional anomalous result was repeated with the addition of colour indicators (methyl orange and bromocrescol green for HCO_3 and phenlphathalein for CO_2).

Testing for nutrients, NO_3, NH_4 and PO_4, was undertaken in the field using standard methods (Table 4.3) of the portable HACH DR/2000 spectrophotometer. (HACH 1989). The benefits of on-site analysis were found to be somewhat negated by an accuracy of $\pm 30\%$

Inorganic sample collection

A sample for cation determination was preserved after filtering through a 0.45 μm filter into a 50 ml plastic screwtop jar, and acidified with 2–3 drops of concentrated HCl. An unfiltered and unacidified sample of at least 200 ml was retained in a triple rinsed HPT plastic bottle for chloride analysis and silica testing. All the above samples were cooled, either in the refridgerator of the mobile laboratory van, or in chilly boxes.

Sampling for ferrous iron was accomplished by transfering a 10 ml aliquot of the fitered sample to a glass vial containing 2 ml of PPDT-DAS reagent, followed by 2 ml of sodium acetate buffer. These were stored out of light contact for laboratory analysis.

Samples were also collected for stable analysis at selected sites. Duplicate samples for δ^{18}O and δ^{2}H analysis were collected in 28 ml glass McCartney bottles, ensuring no head space. The bottles were periodically tightened and stored inverted against the insert rubber seals to ensure an air-tight seal. Samples for δ^{13}C analysis were collected by precipitating all available DIC in a 2 L samples, using saturated $SrCl_2$ under alkaline conditions. This was achieved by addition of 4 ml of freshly prepared 11 M NaOH.

Selected samples were collected for radioistope analysis. Samples for δ^{3}H analysis were collected in 1.0 L glass Schott bottles, taking care that the tops were sufficiently tightened to prevent atmsopheric exchange. Samples for carbon-14 analysis were obtained according to guidelines supplied by the Institute of Geological & Nuclear Sciences (C.Taylor, pers. com., March 2000). Saturated $SrCl_2$ and NaOH solutions were pre-prepared in numbered and pre-weighed 50 ml Schott bottles using anlaytical grade reagents. Samples were collected in accurately pre-weighed and numbered 1 L Schott bottles. The contents of partnered 50 ml Schott bottles was added to each sample bottle, with care that any precipitated impurities were left in the 50 ml bottle. The bottles were then filled to the rim to exclude as much air as possible. Lids on both bottles were periodically tightened.

Rainfall sample collection

A simple rainfall sampler was designed using a modified 50 L plastic drum. A 0.5 m diameter plastic funnel fed into a 1.0 L glass Schott bottle nested within the drum. This design ensured a sample volume of about 1.0 L from a 5 mm rainfall event. To ensure a pure rainfall sample, the funnel was covered by a lid between rainfall events and was fitted with an insect screen mesh. The sampler was located in the township of Leeton, managed by S. Lawson.

The lid was removed prior to an imminent rainfall event and the sample collected

Table 4.3: Summary of water analysis methods

Parameter	Methodology	Concentration Range	DL [a] (mg/L)	Sample Volume (mls)	Holding Period	Lab
HCO_3,CO_2	Titration	-		200	Minutes	Field
NO_3	HACH field kit - Cadmium reduction*	0 to 4.5 as NO_3-N [b]	0.1	50	Hours	Field
NH_4	HACH field kit - Nessler method*	0 to 2.5 as NH_3-N	0.02	50	Hours	Field
PO_4	HACH field kit - Ascorbic acid method*	0 to 2.5	0.01	50	Hours	Field
Fe^{2+}	Colorimetry - PPDT method	~0.05 to 2.5	0.02	10	Weeks	WRL
SiO_3	Colorimetry - Heteropoly blue method	~0.5 to 2	0.1	60	Weeks	WRL
Cl	Agentometric Titration	10 to ¿1000		300	Weeks	WRL
Cations	ICP-AES	-		60	Months	UNSW Geography

[a] DL is detection limit
* HACH 1989
[b] Mid-range test

within 24 hours so as to mimise effects of evaporation and atmospheric exchange on the sample. As soon as practicable, pH and EC were measured. Samples were transferred to duplicate McCartney bottles for δ^{18}O and δ^{2}H analysis, and the remainder into a plastic bottle for subsequent major ion analysis.

4.2.2 Laboratory analysis

A summary of hydrochemical analyses, including method ranges and detection limits is outlined in Table 4.4.

Major cations, trace elements, total P and total S were determined by inductively coupled plasma atomic emission spectroscopy (ICP-AES) at the UNSW Geography laboratory (D.Yu, pers.com.). Samples with high EC were diluted prior to analysis, and the results corrected as necessary. Sulphate was assumed equal to total sulphur (conversion constant $\times 2.994$).

Chloride analysis

Chloride was analysed by the argentometric method described by Beck (1999). In brief, a 50–100 ml sample was titrated with 0.02 M standard silver nitrate ($AgNO_3$), with potassium chromate (K_2CrO_4) solution as an end point indicator. Sample pH was adjusted within the range 6.5–8.5 by addition of sodium bicarbonate. If the EC of the sample was below 500 μS/cm, a 10.0 ml aliquot of standard 0.02 M sodium chloride solution was added prior to titration so as to produce a distinct endpoint, and the necessary corrections made to the result.

Reactive silicon analysis

The heteropoly blue method outlined by Beck (1999) was adopted for analysis of silicate (SiO_3^{2-}), the reactive, soluble form of silicon in water. This method included reduction of reducing species by lead carbonate prior to analysis. Each 20.0 mL water sample was analysed twice, one by addition of molybdate solution, and the other by the addition of deionized water as an individual sample solution blank. Oxalic acid was then added to the mixed solutions, and the absorbance measured by a spectrophotometer at λ of 410 nm.

Iron analysis

Total dissolved iron was determined by ICP-AES (see above). For analysis of Fe^{2+}, the colorimetric method described by Beck (1999) was adopted. This involved addition of the reagent PPDT DAS and sodium acetate buffer to a 10.0 mL filtered (0.45 μm) sample in the field. In the WRL laboratory, the absorbance of standards and samples was measured at λ of 570 nm by spectrophotometer. Fe^{3+} concentration was calculated as the difference between the concentrations of Fe_T and Fe^{2+}.

4.3 Isotope techniques

Conventions for isotope ratios, fractionation and enrichment

As the values are very small, isotope results are reported as a ratio between one isotope to another (eg. $^{18}O/^{16}O$) in the sample and the standard. Conventionally, the ratio is always reported as the heavier isotope to the lighter element. Delta, δ, represents the difference between the sample and standard, and is reported in units of per mil, ‰ (parts per thousand).

$$\delta(\%) = \frac{R_{sample} - R_{standard}}{R_{standard}} \times 1000 \tag{4.3}$$

Where R is isotopic ratio given by the number of atoms for the two isotopes.

The internationally accepted ^{18}O standard for application in hydrology is VSMOW (Vienna Standard Mean Ocean Water), which is artificially prepared and distributed by the IAEA (International Atomic Energy Agency).

$$\delta^{18}O = (\frac{(^{18}O/^{16}O)_{sample}}{(^{18}O/^{16}O)_{reference}} - 1) \times 1000 \text{ ‰ } VSMOW \tag{4.4}$$

Where VSMOW is Vienna Standard Mean Ocean Water.

The standard defined by the IAEA in Vienna for δ^{13}C is VPDB which is considered identical to the original standard PDB. Introduced by Craig (1957), the PDB standard was prepared from the interior of fossil from the Pee Dee Formation in South Carolina. In practice, δ^{13}C is measured relative to the standard NBS-19 which was prepared from a slab of marble and calibrated with the rapidly exhausted supply of PDB.

Fractionation

Isotopic fractionation is a partitioning of isotopes that occurs during physical and chemical processes due to difference in reaction rates. The isotopic fractionation factor, α, describes the partitioning of isotopes between two substances A and B. For example $\alpha = 1.028$ for calcite-water equilibrium indicates that calcite is slightly enriched in δ^{18}O relative to water (Equation 4.5).

$$\alpha_{calcite-water} = \frac{(^{18}O/^{16}O)_{calcite}}{(^{18}O/^{16}O)_{water}} \tag{4.5}$$

Isotopic enrichment, ϵ, is used to express isotopic differences in ‰ notation and is related to the isotopic fractionation factor (Equation 4.6).

$$\epsilon_{X-Y} = (\frac{R_X}{R_Y} - 1) \times 10^3 = (\alpha - 1) \times 10^3 \tag{4.6}$$

The isotopic enrichment for the calcite-water example described above would thus be 28 ‰.

Table 4.4: Potential errors in δ^{13}C sample preparation (data from Bishop 1990).

Procedure	Error magnitude ‰
Filtration time	+1
Oven drying time	+0.4
Precipitation time	-0.1
$SrCl_2$ vs $BaCl_2$ reagents	nil
Polymer vs. glass containers	nil

δ^{18}O and δ^2H analysis

The stable isotopes δ^{18}O and δ^2H were analysed at CSIRO Isotope Analysis Service in Adelaide by standard methods using a Europa Geo 20-20 Duel Inlet Stable Isotope gas Ratio Mass Spectrometer (M. LeFournour, pers.com. 8.8.00).

For δ^{18}O, this entailed preparation of CO_2 gas for spectrometer analysis by overnight shaking of a vacu-tainer with 2 ml of water sample to achieve equilbrium with CO_2 (1 atmosphere pressure, 30^0C).

The δ^2H composition of water samples were determined by circulating the vapour of a 25 μL water sample over uranium metal heated to 800^0C, which reduced hydrogen to hydrogen gas for spectrometer analysis.

δ^{13}C analysis

δ^{13}C of groundwater DIC was analysed relative to the VPDB standard. Field precipitated $SrCO_3$ was filtered and dried at WRL, in preparation for ^{13}C analysis at the CSIRO Centre for Isotope Studies located at North Ryde.

An inert (N_2 atmosphere) glove box was constructed in order to filter samples out of contact with atmospheric CO_2. This method was adopted part way through the sampling program due to contamination concerns raised by the critical appraisal of methods by Bishop (1990) (Table 4.4).

Additional strategies which were adopted to prevent atmospheric contamination included: Dreschsler CO_2 trap fitted to N_2 inlet of the glove box, a slaked lime ($CaOH_2$) trap within the vacuum pump line, and rapid filtering under vacuum (Advantec 47 mm glass fiber filter GL50). The samples were dried within a dessicator.

CO_2 gas samples were generated from $SrCO_3$ by the author by reaction with 103% H_3PO_4 within an evacuated reaction vessel, after equilibration at 25^0C in a water bath. Prepared gases were analysed by Matt Finnigan 252 dual inlet Isotope Ratio Mass Spectrometer in automatic batch mode by B. McDonald (CSIRO Centre for Isotope Studies).

4.3.1 Carbon-14 dating of groundwater

Background

Radiocarbon dating measures the loss of parent radionuclide (eg. ^{14}C) in a sample which decays at a constant rate determined by the half life (eg. 5 730 years for ^{14}C). This means that after 5 730 years, half of the ^{14}C in a given sample will have decayed, and after two half lives, or 11 460 years, half of the remaining ^{14}C atoms will have decayed leaving only 0.25 of the original amount. This process is described by Equation 4.7.

$$t = -8267 \times ln\left(\frac{a_t^{14}C}{a_o^{14}C}\right) \tag{4.7}$$

Where a_t is activity after some time t, and a_0 is initial activity.

Dating by this method assumes that the intial concentration of ^{14}C is known and has remained constant, and secondly, that the system is closed to subsequent gains or losses of ^{14}C except by decay. Details about natural and anthropogenic variations in ^{14}C activity in the atmosphere and correlation of the time scale by independant methods is given by (Clark & Fritz 1997) and will not be elaborated here. Activity of ^{14}C is expressed as per cent modern carbon (pMC) where 100 pMC corresponds to 95% of the ^{14}C concentration of NBS oxalic acid standard (close to activity of wood grown in 1890). Standard pMC results are normalised to a $\delta^{13}C$ = -25 ‰ (Stuiver & Polach 1977).

Initial activity, $a^{14}C$ in groundwater is determined by primarily by exchange of gaseous soil CO_2 with infiltrating water. Once flowing within a confined sytem, ^{14}C activity in groundwater is determined by decay, though interpretation may be complicated by geochemical reaction, diffusive loss and and other factors.

Analysis method

The carbon-14 activity of a limited number of groundwater samples was determined by atomic mass spectrometry (AMS), at the Rafter Radiocarbon Laboratory (Institute for Geological and Nuclear Sciences), New Zealand. The application of AMS enabled a very small (1.0 L) sample to be analysed, in contrast to classical radiometric techniques such as liquid scintillation counting . Total dissolved inorganic carbon was determined by high accuracy gravimetric methods.

Table 4.5: Interpretation of tritium activity in Australian rainwater assuming no mixing (Calf 1988).

Activity TU	Interpretation
< 0.2	Water older than 50 yrs
< 2.0	Water older than 25 yrs
2 - 10	Interpretation difficult, water probably modern
10 -20	Water 15–25 yrs old
$>$20	Probably related to water from peak fall-out period 1960-1964.

4.3.2 Tritium dating of groundwater

Background

Tritium, ^{3}H, is a radioisotope with a half life of 12.43 years. The tritium unit (TU), also known as the tritium ratio (TR), is defined as one tritium atom per 10^{18} atoms of hydrogen. Tritium is produced naturally in small amounts due to the interaction of cosmic radiation with atmospheric oxygen and nitrogen in the troposphere. However, tritium was also produced by thermonuclear explosions in the 1950s and 1960s. In Australia, tritium in rainfall peaked between 50 to 100 TU, but was expected to approach natural background concentrations of between 4–6 TU by the early 1990s (Calf 1987; Allison et al. 1971).

In principle, tritium is an ideal radioistope that is transported conservatively as an integral part of the water molecule unaffected by geochemical reaction. Unfortunately, tritium dating was at the best of times, an ambiguous and non age-specific method (Table 4.5). For example groundwater containing about 2.4 TU could be the result of recharge in 1990, or recharge in 1975. Tritium levels $<$0.25 TU can be attributed to groundwater residence times greater than 50 years, recharged prior to elevated atmospheric tritium levels beginnning in the 1950s. Tritium is now best applied in combination with measurement of the decay daughter element ^{3}He (eg. Plummer et al. 2000).

In this study, some useful insights regarding mixing of younger water in deep aquifers was obtained from tritium analysis at very low levels. This was possible using a technique with a detection limit that is an order of magnitude lower than methods which are more readily available in Australia.

Analysis method

Tritium activity was analysed on selected groundwater samples, at the Institute for Geological and Nuclear Sciences, New Zealand. Very low level detection ($\pm$0.017 TU) was achieved by a high degree of analytical enrichment and state-of-the-art low-level liquid scintillation counting.

4.4 Quality assurance

A range of measures were adopted to ensure the quality of results for water analyses.

Artifacts of sampling were minimised by sampling from purpose built piezometers, with a set purging time or number of bore volumes extracted prior to sampling. Identical equipment and analysis techniques, and the same analyst for chloride and most other species (myself - see laboratory register on CD-ROM) eliminated analysis variabilities as much as was practicable.

Cross contamination of samples was prevented by thorough equipment cleaning procedures in the field and at WRL laboratory. The integrity of RO deionized water and MQ ultrapure water used for rinsing and dilution were checked periodically by ICP-AES analysis.

As discussed above, the quality of data was ensured by the use of standards, and frequent calibration of field meters. A flow cell was used to obtain unstable parameters. Some problems were experienced with the Eh electrode, so suspect data was abandoned and interpretation restricted to generalised trends. Titrations (HCO_3, CO_2, Cl) and nutrient analyses were performed twice, and if not in agreement, analysed a third time. Two duplicate samples were included in each batch of analyses, including those for ICP-AES analysis. A blind duplicate sample was submitted for stable isotope analyses by external laboratories.

Major ion analyses, including those obtained from DLWC archives, were checked for charge balance error (CBE). Electroneutrality is a measure of analytical accuracy since a solution must be electrically neutral. Hem (1992) considered that under optimum conditions, analytical results for natural waters have an accuracy of 5–10%. The CBE of 4 out of 44 DLWC samples exceeded 10%, with a maximum error of 20%. For those samples collected as part of this study, the CBE of 2 of 85 samples exceeded 10%, with a maximum CBE of 12%.

Chapter 5

AQUITARD HYDROGEOLOGY

In this chapter, the hydrogeological importance of aquitards is considered by examing geological controls on vertical leakage, such as permeability of the aquitard matrix and spatial heterogeneity within the aquitard-aquifer system.

5.1 Results

The various units identified in this chapter are referred to as upper, middle and lower silt units (based upon grain size analysis) and as numbered aquifer units. The specific location of these units within the Shepparton or Calivil Formations is considered after the detailed analysis data are presented.

5.1.1 Borehole geophysics

Regional stratigraphic variability is evident in downhole geophysical logs of selected sites on the Lower Murrumbidgee alluvial fan (Figure 5.1). The disimilarity of the stratigraphy at these 5 sites means that it is not possible to differentiate regionally extensive layers and makes assignment of the units identified at the Tubbo site to one or other of the regional formations (Shepparton or Calivil) very difficult.

Geophysical logs for the Tubbo and Gundaline sites, located near the river, are characterised by relatively low bulk EC (100–200 mS/m) and high gamma-ray (150-250 cps) activity within the top 10 m. This contrasts strongly with the high bulk ECa peaks and lower gamma-ray activity at bores further from the river (South Tubbo, CIA Pump and Rice Mill). These differences are interpreted to indicate significant flushing of salt from sediments close to the river and that the recent deposits close to the river represent different sedimentary sources of higher potassium rich (gamma emitting) material. Though shallow sand layers are also indicated by the geophysical logs for South Tubbo and Rice

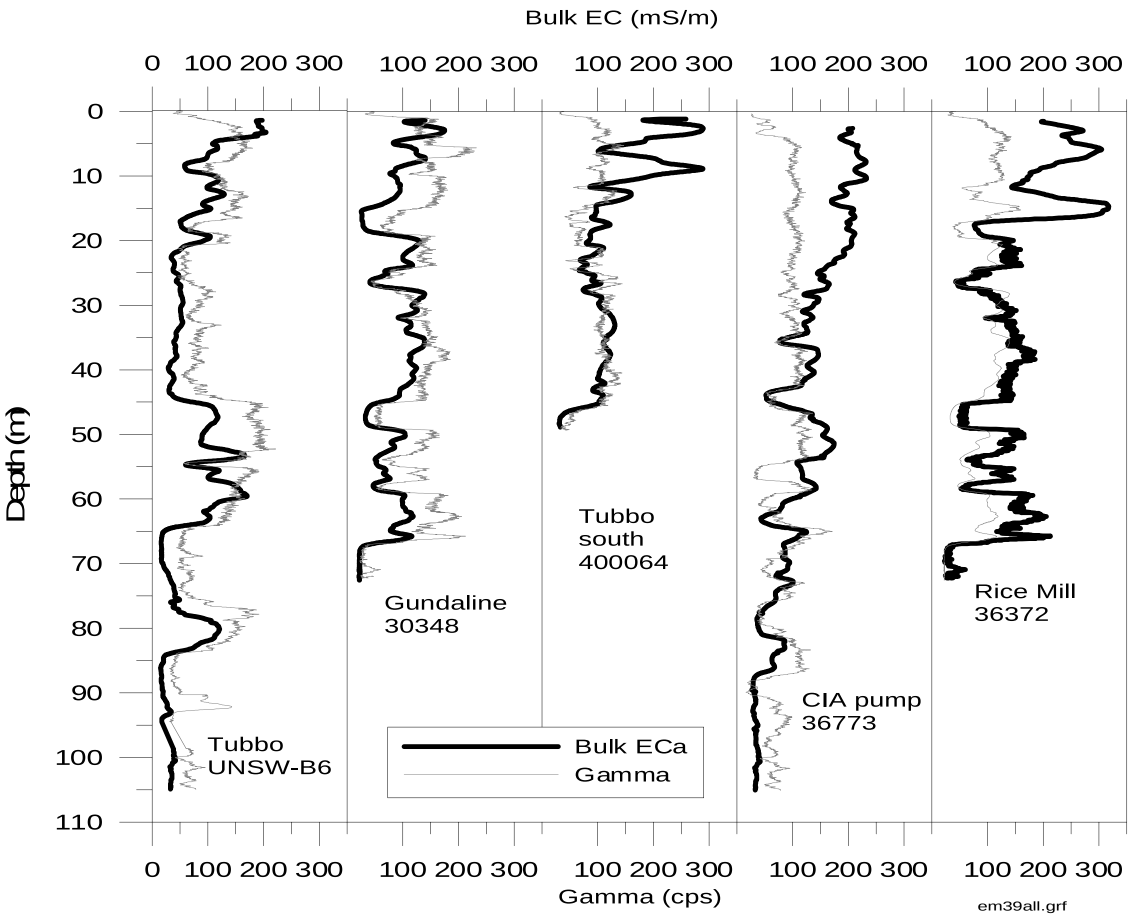


Figure 5.1: Borehole geophysical logs from selected sites in the alluvial fan showing regional variability. (Source for Tubbo south and CIA pump site logs: R.I. Acworth, pers. com.)

Mill bores, bulk EC exceeds 250 mS/m, suggesting that downwards flushing of salts has been minimal at these sites. A relatively constant gamma response to 35 m depth at the CIA site is associated with a downwards trend of bulk EC.

Although silt units occur at similar depth in all three bores at the Tubbo site (Figure 1.4), direct correlation is not possible, even over a distance of 40 m between 59113 and UNSW B6. At a distance of 1 770 m from the massive silt units of UNSW B6 and 59913, multiple layers less than 5 m thick form the middle and lower silt units.

Geophysical techniques proved useful in delineating clayey and sandy deposits, and also variability within specific lithological units. The results of the gamma-ray and bulk electrical conductivity logging for bore UNSW-B6 are shown in Figure 5.2. Five distinct units are present although there is also considerable heterogeneity.

Upper silt unit - This silt unit extends from the surface to a depth of 21 m. Sand lenses occur at approximately 8 m and 18 m, indicated by lower bulk conductivities and lower gamma-ray activities. The highest values of bulk conductivity (200 mS/m) occur in the top 4 m. The gamma log peaks at approximately 150 cps in this unit. Bulk electrical conductivity peaks at 200 mS/m in the top 5 m but then falls to background levels of 50–60 mS/m in the sand units and approximately 125 mS/m in the silt units.

Aquifer 1 - The top sand unit extends between 21 m and 26 m and is clearly indicated by low bulk conductivity (approximately 40 mS/m) and lower gamma-ray activity (50 to 100 cps).

A indurated clayey sand (26–32 m) is seen in the physical analysis data for the core but is not seen as a significant feature in the bulk electrical conductivity log, though a small increase in gamma-ray activity was observed.

Aquifer 2 - This unit extends from 32 m to 45 m and is indicated by uniformly low bulk electrical conductivity (40–50 mS/m) and lower gamma-ray activity. The gamma-ray activity is significantly higher than the major sand units in other parts of the log (21–26 m; 65–75 m and 84–90 m).

Middle silt unit - The middle silt unit extends between 45 m and 64 m. It is clearly indicated by a uniform band of unusually high gamma-ray activity readings (200 cps) between 45 m and 53 m. A thin sand lens separates the upper massive clay from lower gamma activity clay at the bottom of the unit. Bulk conductivity values of approximately 150 mS/m occur through the unit but oscillate between a minimum of 90 mS/m and a maximum of 170 mS/m.

Aquifer 3 - This sand unit clearly extends between 64 m and 75 m. Low values of both bulk conductivity (25 mS/m) and gamma-ray activity (55 cps) indicate a sand.

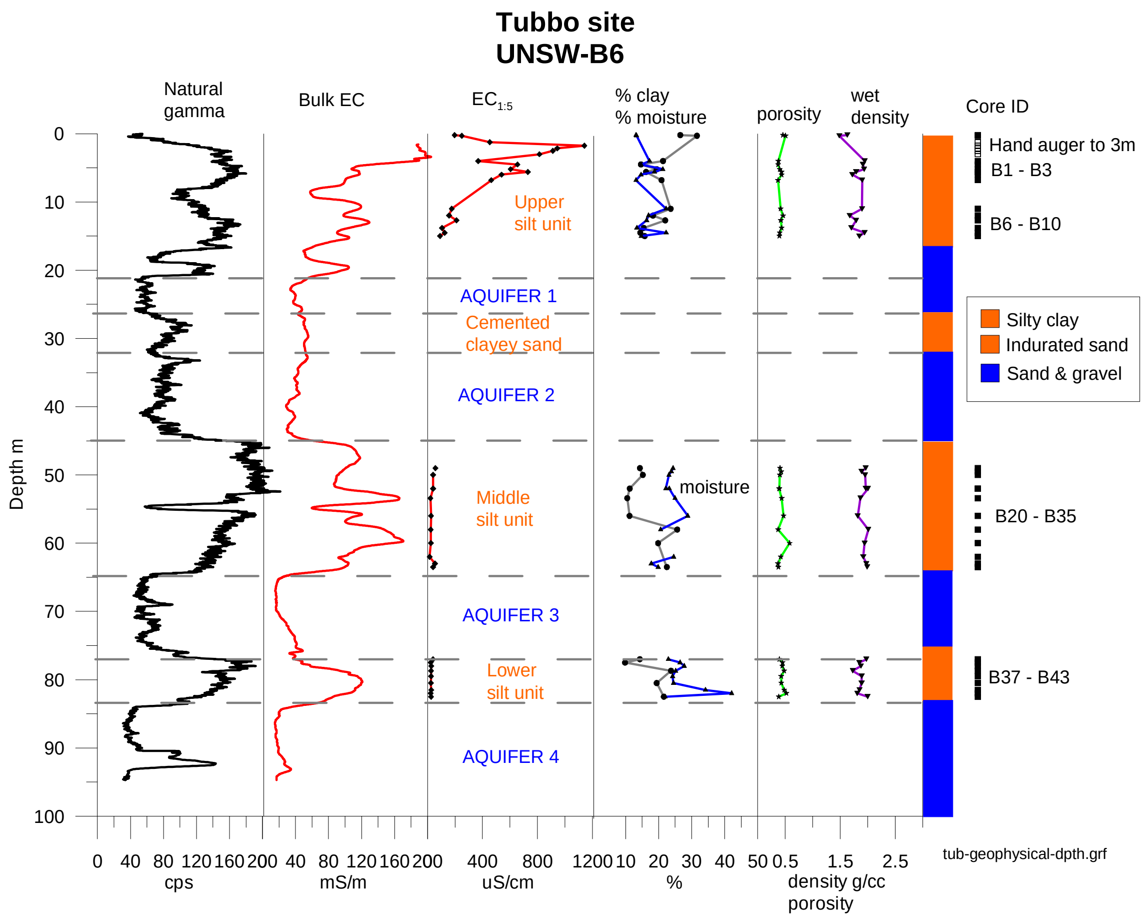


Figure 5.2: Geophysical logs and physical parameters derived from core analysis for UNSW-B6.

Lower silt unit - The lower silt unit extends between 75 m and 83 m and is indicated by a sudden increase in gamma-ray activity and bulk conductivity.

Aquifer 4 - The lower sand unit extends from 84 m to the base of the logged hole.

5.1.2 Electrical images

Considerable variation in apparent electrical conductivity (ECa) is evident in the electrical images presented in Figures 5.3 and 5.4. In electrical images from the Tubbo site, two areas of high ECa which are approximately 80 to 100 m wide, are separated by a relatively low ECa feature which is about 40 m wide. The high ECa area on the irrigation channel side extends to approximately 15 m depth, correlating well with the Upper silt unit of piezometer UNSW-B6. The low ECa feature appears to trend in a north-easterly direction, since the two electrical images are parallel but the second offset towards the north by about 50 m (Figure 1.3). At the DLWC bores, 1.8 km away (Figure 5.4A), the surface layer is thinner but more extensive.

Bulk ECa is significantly higher, and more uniform (Figure 5.3) at the Rice Mill site, which is located north of the CIA (Figure 1.2).

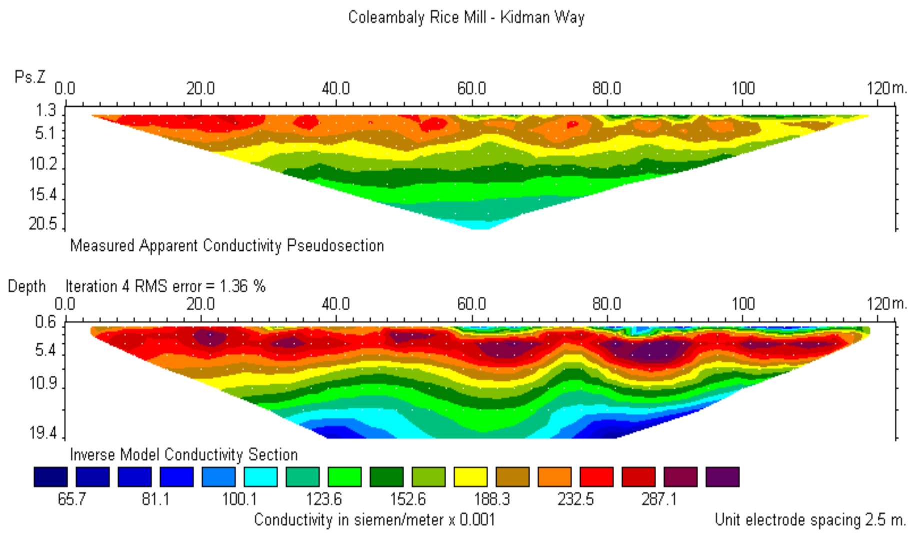


Figure 5.3: Electrical image for Rice Mill site. Upper image - Measured resistivity section and Lower image - Inverse model of true resistivity.

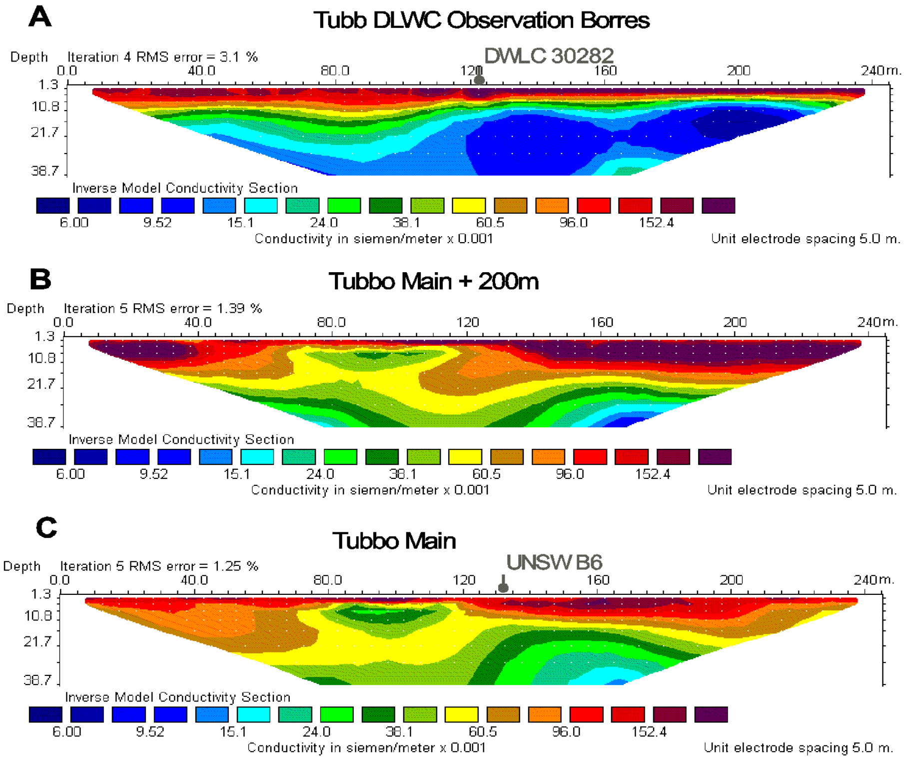


Figure 5.4: Electrical images (inverse models only) showing palaeodrainage features in upper silt unit. **A** EI-3 through DLWC 30282 and **B** EI-2 located 200 m east and parallel of EI-1, with a 50m offset to the north and **C** EI-1 through nested piezometer site UNSW-B6 (Source for EI-1 and EI-3: R.I. Acworth, pers. com.)

5.1.3 Lithological sections

Despite some uncertainty associated with sparse deep bore data on the Tubbo irrigation property, three main clayey silt units, recorded as 'clay' on driller's logs for existing bores, are evident (Figure 1.4). A shallow aquitard between 7.6 and 16.6 m thick is present at all three bore locations (59113, B6 and 30282). The middle aquitard between 50 and 64 m depth is massive in the west and in the east occurs as three thin (2 to 4.5 m thick) clay strata, interbedded by sandy gravel. A deep aquitard is found between 70 and 86 m depth, thinning from west (14 m thick) to east (two layers ∼5 m thick).

Clay-rich deposits form a significant proportion of the alluvial profile with 27, 36 and 38% at bores DLWC-30282, UNSW-B6 and GW59113 respectively. Although the depth to bedrock at these bore sites is unknown, weathered granite at a depth of 120 m is recorded for nearby observation bores (Table 5.1).

Table 5.1: Basement geology for bores where data is available (see Figure 1.2 for locations).

Location	Bore	Depth (m)	Base alluvium	Bedrock
DP east	30489	120.4		w.granite [1]
Rice Mill	36372	162-188	carbonaceous clay	
		200-214		w.granite
CIA pump	36773	178-192	carbonaceous	granite
	36774	190-192		conglomerate, porphory
Gundaline	30348	140.2	carbonaceous clay	
Tubbo	59113	128	carbonaceous clay	
North of Tubbo	400029	132-148	carbonaceous	coal
Northern CIA	400064	150	sandy	coal
Northern CIA	400066	123	carbonaceous	coal

1 w. weathered

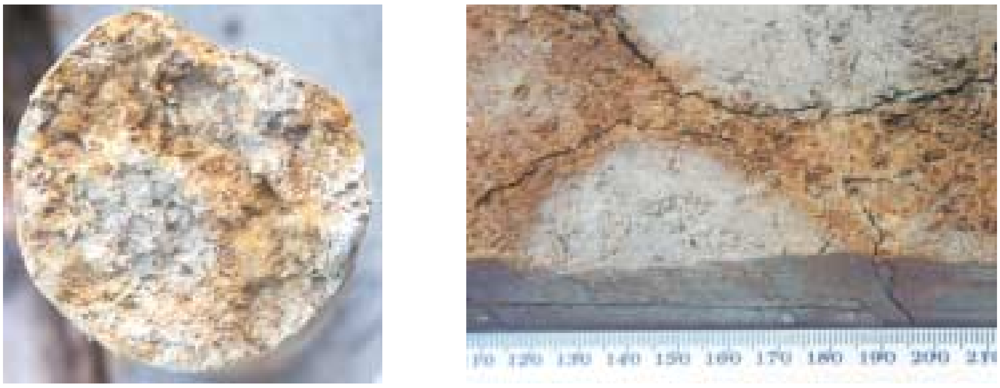
5.1.4 Physical description

Detailed physical attributes of each of the aquitards were described by Timms & Acworth (2002b), the highlights of which are outlined here, together with new mineralogical data for the indurated clayey sand (Table 5.2 and 5.3).

Visual examination of the cores revealled significant heteregeneity, which was attributed in part, as the effects of weathering (Figure 5.5). Microfractures surrounded by 1–2 mm of orange oxidised clay were rare in core from the deep aquitard. The clayey sand encountered between 26 and 32 m depth was sufficiently hard to require bit cutting since the core barrel could not be pushed through the sand. The strength of this layer was estimated to be equivalent to a soft coffee rock (M. Groskops, pers. com.). The hard clayey sand layer was thought to be cemented. However, the lack of effervescence upon application of dilute acid, indicates that carbonate was not present. The matrix of the clayey sand disintegrated during drying, and was not sufficiently hard to be silcrete.

The modal grain size was silt, with % clay for the shallow aquitard ranging between 14.4 and 31.6%, the middle aquitard between 10.5 and 25% and the deep aquitard between 9.8 and 23.7%. Figure 5.2 shows the variation of % clay with depth, and shows the increasing clay content towards the ground surface. Average moisture content varied between 16.6 and 27.7%.

Distinctive clay mineralogy was observed for each of the aquitards. The clay fraction of the deep aquitard was comprised of 66% kaolinite, and the middle and shallow units were dominated by kaolinite-illite and illite suites respectively. The highest proportion of



A B

Figure 5.5: Photographs of heterogeneities in silt core **A** Cross-section **B** Grey kaolinite relicts in weathered mottled orangey clay (large fractures developed during core storage).

Table 5.2: Bulk mineralogy of the indurated clayey sand at Tubbo (see Timms & Acworth 2002b, for mineralogy of clayey silt units).

Depth (m)	Qtz	Flds	Ant	Gyp	Clc	Pyr	Gth	Mgn	Hmt	Mordenite	Siderite
31	M	M	-	-	-	-	-	-	?T	?T	?T
31	M	M	-	-	-	?T	-	-	?T	-	-

D = dominant (>60%), A = abundant (60-40%), M = moderate (40-20%), S = small (20-5%), T = traces (<5%)
Qtz = quartz, Flds = feldspar, Ant = anatase, Gyp = gypsum, Clc = calcite, Pyr = pyrite, Gth = goethite, Mgn = magnetite, Hmt = haematite

Table 5.3: Clay mineralogy of bulk and clay fraction for indurated clayey sand at Tubbo (see Timms & Acworth 2002b for mineralogy of clayey silt units).

Depth (m)	Bulk			Clay				
	Kaolinite	Illite	Smectite	Kaolinite	Illite	Smectite	Chlorite	K/I ratio*
31	M	M	S	45	33	22	-	6.2
31.5	M	M	S	47	36	23	-	4.9

*glycolated <2mm fraction

A B

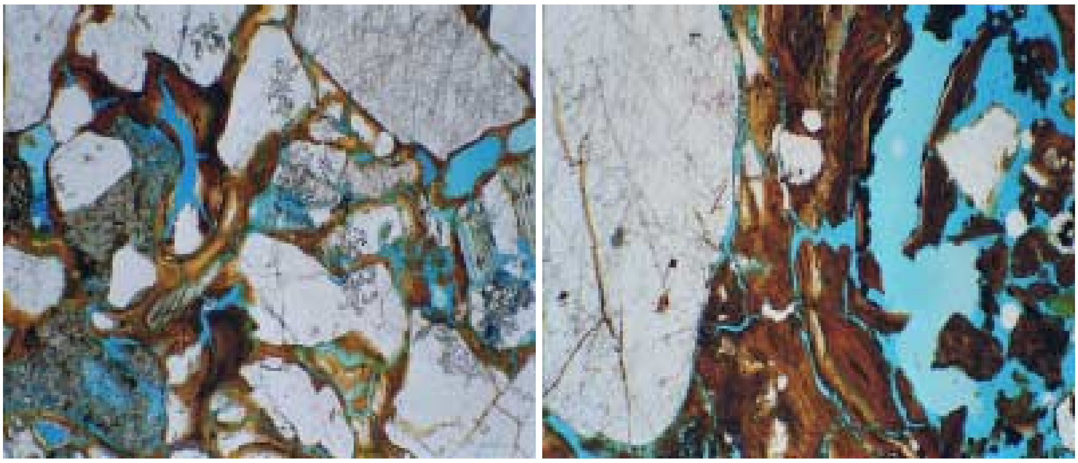


Figure 5.6: Thin-section microphotographs of indurated clayey sand with pore spaces stained blue (x50 magnification, field of view 4 mm, plain polarised light) **A** Clay matrix supported quartz and lithic fragments (No. B16A/2) **B** Remnants (right) of opaque mineral (pyrite ?) which appears to be the source of Fe staining of clay matrix, large quartz grain (left) (No. B14/1).

smectite found was 38% at a depth of 52.7 m within the middle silt unit. However, the proportion of smectite clay increased towards the surface of the upper silt unit.

In contrast to the upper and middle silt units, the indurated clayey sand was dominated by kaolinitic clay (Table 5.3), and contained traces of haematite, mordinite and siderite (Table 5.2).

Total subsurface salt storage contained within clayey silt (35 m total thickness) was about 11.8 kg/m^2. Of this salt store, the upper silt unit accounted for 86%, or the equivalent to 102 tonnes/ha of salts within 15 m of the ground surface, if salt laden silt was distributed homogeneously over this distance.

5.1.5 Thin section analysis of indurated clayey sand

The nature of the indurated clayey sand (26–32 m), delineated in Figure 5.2, was assessed by thin sections prepared and analysed by the methods outlined in Chapter 4.1.4.

On a macroscopic scale, visual examination of thin sections revealled significant heterogeneity. Reddy-brown iron staining was evident parallel to bedding, along with other small-scale bedding and cross bedding features. There were also fractures evident that were oriented at an acute angle to bedding.

Examination by petrographic microscope revealled that quartz grains were the dominant constituent, making up nearly half of the sand grains (Figure 5.6). A fine grained matrix was found to infill much of the pore space, accounting for 23.3% of the total point-count, compared with free pore space of 16%. Lithic fragments were common, and minor quantities of biotite and other minerals such as an opque species, possibly pyrite.

Pore throat diameters were found to vary from zero (totally infilled by fine grained matrix) to 293 μm (about 0.3 mm), with median diameters of 18.3 and 24.4 μm (N=50).

Tables which detail the results of microscope thin section analysis, including pore throat diameters and porespace versus mineralogy, are available on the accompanying CD-ROM.

5.1.6 Falling head permeameter testing of core samples

Over 30 falling head permeability tests were conducted on core samples from the middle silt unit (4) and indurated clayey sand (4), according to the methods outlined in Chapter 4.1.3. Results are summarised in Table 5.4.

The first few tests were run with WRL tap water (EC = 150 μS/cm), and also Tubbo site groundwater from the Upper Shepparton aquifer (B1 EC = 972 μS/cm) and the Lower Shepparton-Calivil aquifer (B4 EC = 160 μS/cm). Measured K_v for cores tested with insitu groundwater tended to be 2–4 times higher than those tested with WRL tap water (eg. compare tests 7&11 with 8&12). Since both waters were fresh and of approximately the same EC, this effect was attributed to likely differences in major ion type that resulted in cation exchange with the clay.

An order of magnitude lower K_v were measured for cores in larger 11.3 cm diameter chambers compared with small diameter chambers of 7.4 cm diameter (eg. compare tests 10 & 13 with 8 & 12). If scale were to be important, it would be expected that the larger the volume tested, the higher the permeability due to heterogeneity such as micro-fracturing. This was not the case however, at least for small change in volume concerned. It is possible that edge effects of preferential flow were responsible, since there was a higher proportion of edge per volume for the smaller diameter chambers.

Variable K_v during individual tests was observed, contrary to Darcy's Law (eg. Figure 5.7). This could be due to viscosity changes of the permeant, and swelling or consolidation of the core sample in response to changing pressure gradient, or chemical reaction with the permeant.

Correction for viscosity resulted in a more constant K_v, particularly for those tests which ran over more than a few minutes. It was also possible that swelling resulted in increased permeability, as the top of the cores were unconfined. However, swelling of only a few millimeters was expected based on the results of consolidation testing. Swelling of 0.5

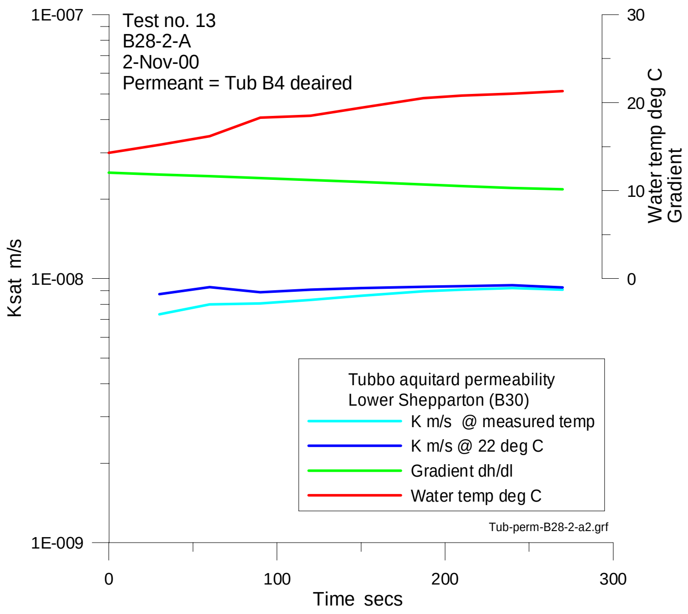


Figure 5.7: Example of permeability variation due to gradient and temperature changes during a falling head permeameter test.

cm for instance, would be expected to increase the porosity from 0.42 to 0.52, an increase of 20%. Given the log-normal relationship between hydraulic conductivity and porosity, it is probable that such a change in porosity would increase permeability by about an order of magnitude (Neuzil 1994).

It should also be noted that swelling may also have occurred prior to testing due to lower effective stress as the cores were extracted from the ground and during subsequent storage at atmospheric pressure. Without detailed laboratory and field measurements of core parameters it is not possible to quantify this artifact.

Chemical reaction between the clay and the permeant may also cause varied K_v during tests, and between repeated tests. For example, flushing with sodic water may cause dispersion of the clay and decreased permeability. This appears to be the case for the clayey sand which generally showed decreased K_v both during and between repeated tests. This may be attributed to cation exchange of sodium which changes soil structure. Hydraulic conductivity of soils is known to decrease with increasing SAR of leaching water (Appelo & Postma 1996).

Increased K_v between tests, that occurred for the lower permeability silty clay may be attributed to increasing degree of saturation. It is possible that saturation in a water bath prior to testing was not completely effective.

Table 5.4: Summary of permeability test results for core samples.

Test no	Core ID	Sample depth (m)	FluidID	Saturation time	Final K_v (m/s)	Final K^v @22 degC (m/s)	K change during test (m/s)
Middle silt unit							
5*	B30-1-b	59	WRL tap**	1 week	1.2E-09	1.3E-09	-3.1E-10
6*	B30-1-c	59	WRL tap**	1 week	1.8E-08	1.8E-09	-2.4E-09
7	B28-2-A	57.8 - 57.9	Tub B4	1 month	2.3E-08	2.4E-08	-1.1E-08
11	B28-2-A	57.8 - 57.9	Tub B4	1 month	9.1E-09	9.3E-09	5.2E-10
10*	B28-3-D	57.6-57.7	WRL tap	1 month	9.6E-10	9.8E-10	4.8E-10
13*	B28-3-D	57.6-57.7	WRL tap	1 month	1.2E-09	1.3E-09	1.0E-10
8	B28-4-B	57.4-57.5	WRL tap	1 month	9.6E-09	9.2E-09	4.2E-09
12	B28-4-B	57.4-57.5	WRL tap	1 month	6.2E-09	6.4E-09	9.0E-09
18	B30-1-d	59	WRL tap	months	1.2E-08	1.3E-08	7.0E-10
Indurated sand aquitard							
19	CS-6-1	31.8	Tub B1	1 hour	1.7E-07	2.0E-07	-1.9E-07
22	CS-6-2		Tub B1	1 hour	1.1E-07	1.4E-07	-2.4E-08
20	CS-4-1	29	Tub B1	1 hour	1.4E-07	1.7E-07	2.6E-08
21	CS-4-2		Tub B1	1 hour	9.7E-09	1.2E-07	-6.9E-09
23	CS-8-1	27.8-27.9	Tub B1	1 hour	1.3E-06	1.6E-06	-1.9E-07
25	CS-8-2		Tub B1	1 hour	1.2E-06	1.3E-06	-2.5E-06
24	CS-7-1	27.9-28.4	Tub B1	1 hour	2.3E-07	2.7E-07	-2.4E-07
26	CS-7-2		Tub B1	1 hour	1.5E-07	1.8E-07	1.1E-07

* Large diameter chamber
** not deaired

After taking these various experimental factors into account, the K_v of the clayey sand was found to be between 10^{-6} and 10^{-7} m/s. Permeabilty of the middle silt unit was considerably lower, at between 10^{-9} and 10^{-10} m/s.

5.1.7 Oedometer testing of core samples

Oedometer testing was conducted according to methodology outlined in Chapter 4.1.3. Detailed test results are available on the accompanying CD-ROM. Design of the consolidation test, depended on calculated total stress, from which effective stress could be calculated, given known pore pressures (Figure 5.8). Total normal stresses of 400, 800 and 1600 kPa were applied to the core subsamples causing settlement of up to 1 mm within a

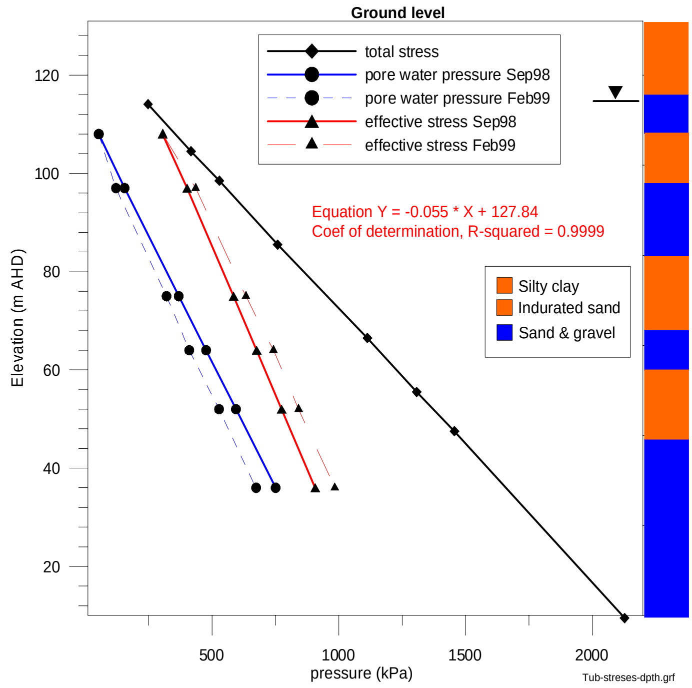


Figure 5.8: Pore pressure variation with depth at Tubbo site, with calculated total and effective stresses.

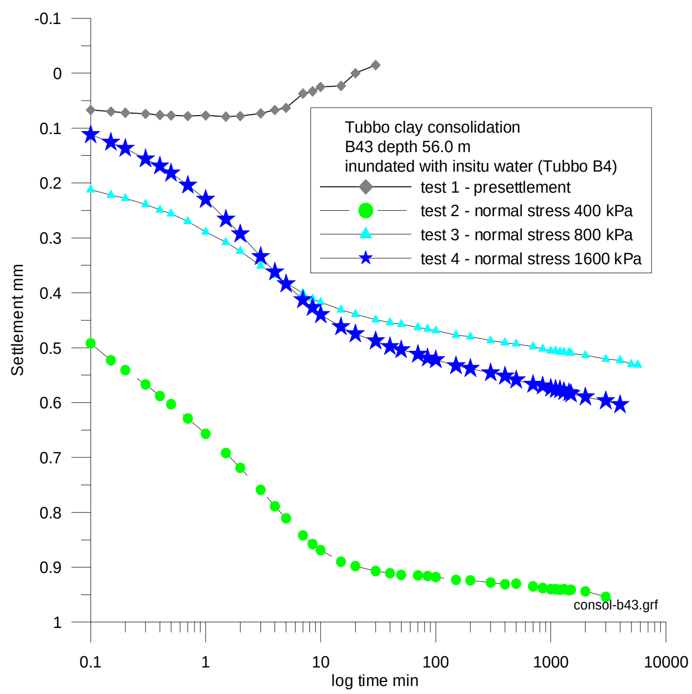


Figure 5.9: Example of consolidation test where settlement occurs due to dewatering of the aquitard matrix with increasing normal stress.

Table 5.5: Measured c_v values compared with typical values for aquitard units.

Lithology	c_v (m^2/year)	Reference
Middle silt unit	2.27	Experimental
Lower silt unit	1.7–3.4	Experimental
Boston Blue clay	12 ±6	Holtz and Kovacs, 1981, p404
Chicago silty clay	2.7	"
Louisanna clay (fluvial - illite)	0.3	Lambe, 1951, p84
Mexico City clay	0.3–0.5	"

2 cm thick sample (Figure 5.9).

The coefficient of consolidation, c_v, and coefficient of compressibility, a_v, were then determined by graphical techniques to be within the range expected for silty clay (Table 5.5). Then, together with moisture content and porosity data, the coefficient of volume change, m_v (Equation 5.1) and hydraulic parameters K_v and S_s were derived (Equation 5.2). The results reflect the hardened nature of the deposits (Table 5.6).

These measured geotechnical parameters formed the basis of estimates for specific storage and storativity appropriate for in-situ stresses (Table 5.7).

$$m_v = \frac{a_v}{(1 + e_0)} \tag{5.1}$$

Where m_v is coefficient of volume change (/kPa), a_v is coefficient of compressibility (/kPa), and e_0 is initial void ratio.

$$c_v = \frac{K_v(1 + e_0)}{\rho_w g a_v} = \frac{K}{9.8 m_v} \tag{5.2}$$

Where c_v is the coefficient of consolidation (m^2/day), ρ_w is density of water, a_v is coefficient of compressibility (1/kPa) and K_v is hydraulic conductivity (m/day).

5.1.8 Specific yield and effective porosity

Specific yield, S_y, is the ratio of the volume of water that drains from a saturated material under gravitational force, relative to the total volume of the rock (Fetter 1994). Conversely, specific retention, S_r, is that proportion of the porosity, ϕ, which does not drain (Equation 5.3).

Table 5.6: Measured m_v values compared with typical values for aquitard units.

Lithology	State	m_v (m^2/MN)	Reference
Alluvial clay	Soft	> 1.0	Waltham, 1994
	Firm	0.3 - 1.0	"
Till & Tertiary clays	Stiff	0.1 - 0.3	"
	Very stiff	0.05 - 0.1	"
Lower silt unit		0.031	Experimental
	Hard	< 0.005	
Middle silt unit		0.003	Experimental
Rock	Very hard	0.0001	Freeze and Cherry, 1979

Table 5.7: Summary of derived hydraulic parameters for aquitard units.

Test	Stress range (kPa)	K (m/s)	Specific Storage (1/m)	Aquitard thickness (m)	Storativity S
Middle silt unit					
B233	400-800	4.50E-12	3.40E-05	19	6.50E-04
Lower silt unit					
B40b3	400-800	1.9E-11	3.5E-04	8	2.8E-03
B43t3	400-800	2.6E-11	1.1E-04	8	8.5E-04
B403	400-800	3.2E-11	3.0E-04	8	2.4E-03

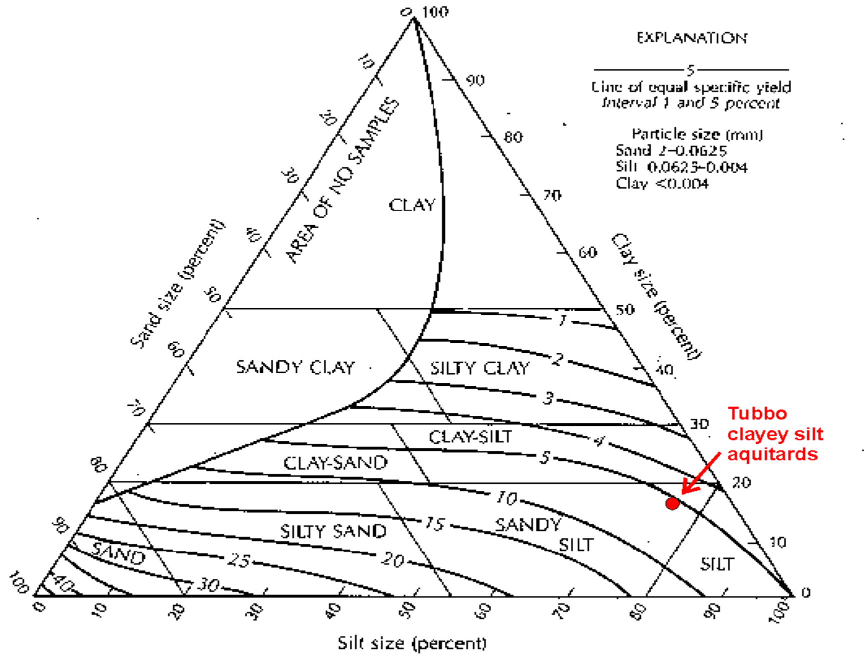


Figure 5.10: Specific yield estimate based on texture (Johnson 1967).

$$\phi = S_y + S_r \tag{5.3}$$

The S_y for the clayey aquitards were estimated from textural analysis data using the classification of Johnson (1967) (Figure 5.10). For these units, characterised by a clay content of 23%, and silt content of 76%, the S_y was about 5%. For the indurated clayey sand S_y was estimated to be 9% according to the following equation (Younger 1992):

$$S_y = 0.54\phi \tag{5.4}$$

Such a low specific yield means that this clayey sand has a very large specific retention, which can be attributed to the large proportion of isolated pore space. This is consistent with values reported by Fetter (1994) of 3–12 % for a sandy clay.

Porosity measurements were used to estimate K according to the method of Younger (1992). For indurated sandstones, K was found to be proportional to porosity (ϕ) by the following equation:

$$K = A\phi^B \tag{5.5}$$

Where A and B are coefficients

An indurated sand deposit with an average ϕ of 0.16 was thus estimated to have a K of 1.6×10^{-6} to 8.6×10^{-6} m/s. The range is due to coefficients of A and B varying from fine to coarse sand.

5.1.9 Slug test of lower silt unit

Three compressed air slug tests were conducted on piezometer B5, so as to determine permeability of the lower silt unit on a field scale.

Water levels in piezometer B5 recovered from a 40 kPa air slug in about 3 minutes. The hydraulic conductivity, derived by the Hvorslev method (Freeze & Cherry 1979; Kruseman & de Ridder 1990), was determined to be 2.9×10^{-5} – 4.3×10^{-5} m/s. Further details are available from the slug test hydrograph and Hvorslev interpretation supplied on the accompanying CD-ROM.

Such rapid recovery water level recovery resulted in a relatively high permeability, 5 orders of magnitude greater than permeability determined on core sub-samples by falling head permeameter. This difference could be due to several factors, including fracturing which increases field scale permeability.

However, this result is more likely to be an artifact of slug testing due to the small volume of water which was displaced. Slug testing equipment with a maximum operational pressure of 40 kP, and a 10 m head pressure transducer were used. Consequently, <20 L of water were displaced which because of the 3 m long piezometer screen, and gravel pack, would not have significantly penetrated the formation.

5.2 Aquitard permeability

5.2.1 Evidence for leakage through aquitard matrix.

Timms & Acworth (2002b) reported that about 10.2 kg/m^2 of total salts is stored within 15 m of the surface, which represented 91% of the total subsurface aquitard storage. This quantity is equivalent to 102 tonnes/ha of salts if salt laden silt was distributed homogeneously over this distance. Only 4.38 kg/m^2 (37%) of salt is found within the 3 m of the surface, which is the typical maximum depth of soil surveys. The very low (10–100 mg/L) total soluble salt content which remains in the middle and lower silt units suggests that leakage has occured through the aquitard matrix over the long term.

In the mid-Tertiary Lower Geera Clay, by comparison, leaching resulted in a total soluble salt decrease from 40 000 mg/L to about 10 mg/L (Jones et al. 1994). Several volumes of pore water are likely to have flushed these units since deposition several million years before present.

5.2.2 Related to consolidation state

The permeability of aquitards may be reduced over time, as pumping results in dewatering and consolidation of the aquitard matrix. Subparallel reorganisation of clay particles resulting from physical compaction is reflected by porosity (or void ratio), density, hardness and permeability (Chamley 1989). Changes over time reflect processes such as dessication, post depositional weathering and consolidation, and the stress history of the deposit.

The degree of consolidation does not appear to increase with depth, since porosity is a constant 0.42 (or void ratio 0.72). Porosity is within the range of 0.4 to 0.5 porosity reported for the Geera Clay at 150–180 m depth (Brown & Radke 1989). This may reflect the fact that the sediments are effectively fully consolidated. It is possible, however, that constant porosity may be due to unloading and expansion of clay cores once large effective in-situ stresses were removed during core recovery and storage.

Normally consolidated deposits were defined by Skempton (1970) as those never subjected to greater pressures than modern conditions. Aquitards in the Murrumbidgee area are not expected to have been overconsolidated by tectonic events or major depositional loading or erosional unloading. Slight overconsolidation is likely as a result of intermittent deposition and cyclical dessication of flood plain alluvium. The high density of floodplain clay at 0.6 m depth reported by Hornbuckle & Christen (1999) which is equivalent to the lower end of the density range in the aquitards suggest that dessication is a significant factor for these 'very hard muds'.

Post-deposition weathering can also result in stiff to hard overconsolidated silts and clays (Firman 1994). Clay from the middle silt unit is hard and within the lower silt unit is very stiff to hard (Table 5.6). The high wet density of 1.5–2 g/cm^3 (Figure 5.2) is consistent with a bulk density of 1.56 g/cm^3 at 0.6 m depth for Yooroobla clay on the Murrumbidgee flood plain (Hornbuckle & Christen 1999).

The fact that the younger, shallower aquitard is an order of magnitude harder than the deeper aquitard may be a function of mineralogy. The mineralogy of the Lower Shepparton aquitard is dominated by illite, which is efficiently compacted (Rowe et al. 1995), in contrast with koalinite which is the dominant mineral of the Calivil aquitard. All other factors being equal, a well consolidated illitic aquitard would be expected to have a hydraulic conductivity 10–100 times less than a kaolinitic aquitard.

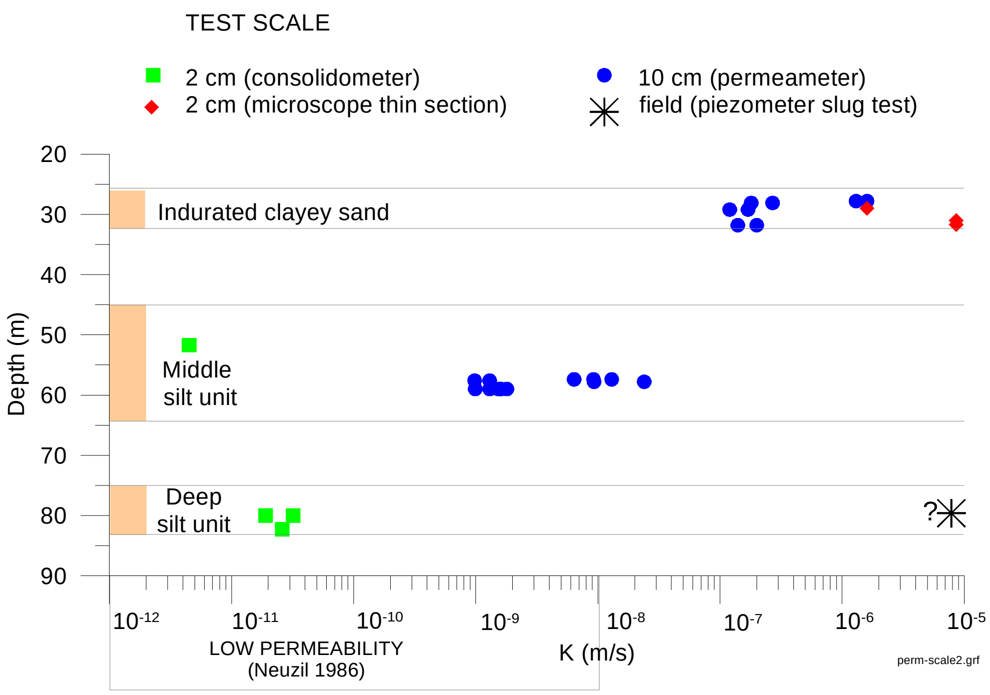


Figure 5.11: Scale dependent hydraulic conductivity for the indurated clayey sand, Lower Shepparton and Calivil aquitards.

5.2.3 Scale dependency

Based on available information from 4 different methods of investigation, permeability was scale dependent within each of the saturated silty units at the Tubbo site. Numerical flow modelling, to be discussed in the next chapter, provided independent permeability estimates for the laterally extensive indurated clayey sand. Here, data from at least 2 methods for each aquitard was compared (Figure 5.11). There was less confidence in the values derived from microscope thin section, and slug test, then the results from oedometer and permeameter testing.

According to results from falling head permeater testing, the indurated clayey sand had the highest permeability, and K_v varied over 1 order of magnitude (10^{-7} m/s). The lowest permeability unit was the middle silt unit, with permeater results of about 10^{-8} m/s, $>$ 3 orders of magnitude greater than consolidometer results. The scale dependence of permeability is common with many other studies of clay systems reviewed in Chapter 3.1. It appears that scale dependence remained important at depth, in contrast to a site in Saskatchewan reported by van der Kamp (2001) where permeability by different methods were similar below the oxidised zone.

K_h determined by slug testing for the lower silt unit (B5) was 5 orders of magnitude greater than permeability determined by consolidation tests, although permeameter results, if tested, would likely be only about 3 orders of magnitude less. This may be attributed to several factors. Slug tests measure K_v, while laboratory methods measure K_h and bedding laminations are likely to cause anisotropy of at least 2 orders of magnitude. There is also the possibility that the piezometer screen intersects small sand deposits or fractures. Such an effect was noted by Keller et al. (1989). However, this single K value should be viewed with caution since for reasons discussed in Chapter 5.1.9, permeability measured by slug testing may be more a function of piezometer installation and development, rather than the silt unit itself.

Evidence for fracturing

There is some direct evidence that the deep silt unit, at least is fractured. Localised oxidation of grey reduced clay along microfractures may be a post-depositional feature associated with groundwater pumping from the underlying sand and gravel aquifer. Preferrential flow may also occur through mottled zones in the upper and middle silt units and minor fractures in the deep silt unit.

5.3 Aquitard windows

Significant lateral variation of apparent electrical conductivity (ECa) is evident in the electrical image sections, EI-1, EI-2 and EI-3 (Figure 5.4). Variation in electrical conductivity is the result of a combination of factors including porosity, saturation, clay content, and the total dissolved salts in the saturating fluid. In saturated homogeneous soils, the electrical image provides an indication of salinity variation (Acworth & Jankowski 1997; Acworth & Beasley 1998). At the Tubbo site, where the water table is at approximately 20 m depth, it is likely that the lateral changes in ECa are dominated by variation in clay content and associated soluble salts.

Stratigraphic architecture inferred from the electrical images is consistent with high ECa areas typical of interchannel clayey deposits and low ECa areas indicating palaeodrainage features. The northeasterly trend of a palaeodrainage feature inferred from EI-1 and EI-2 is consistent with the trend of surface palaeodrainage features (Figure 2.2). A palaeodrainage feature at more than 20 m depth is evident at EI-1 (lower right of image) and was also identified as a sand layer in borehole UNSW-B6. It is possible that this palaeodrainage feature has migrated upwards and laterally over time to the near surface position (centre left of image). A thin veneer of high ECa material covers this inferred palaeodrainage feature.

The palaeodrainage distribution of van Dijk & Talsma (1964) was based on the occurrence of channel sands as outcrop. The outcrop would represent a more complex three dimensional palaeodrainage system extending to bedrock. palaeodrainage channels would appear in plan view as a beaded belt, embedded in a three dimensional, though non- laterally extensive, matrix of clay (Galloway & Hobday 1996). Coarse channel fill sediments, in this form of river deposit, would account for only between 5 and 20% of the total volume. Weber & van Geuns (1990) estimated that at least 2–4 bores per km^2 are required to adequately describe stratigraphic architecture of mixed-load channel deposits.

The presence of obscured palaeodrainage features in an area regionally and previously defined (van Dijk & Talsma 1964) as an inter-channel area, is important when the significance of leakage from irrigation is assessed. There is only a thin layer of clay covering the sands. Dessication cracking may result in rapid by-pass flow carrying pesticides, herbicides and fertiliser into the underlying partially saturated aquifer system. Clearly, the shallow aquifer is more vulnerable to contamination by agricultural chemicals than if covered by a continuous clay layer as assumed by Bowmer et al. (1998). Additional electrical image sections and lithological data, are required to develop a clearer picture of the spatial distribution of clayey material on a scale suitable for paddock scale management.

5.3.1 Extent of the indurated clayey sand

The lateral extent of the indurated iron-rich clayey sand on the Lower Murrumbidgee alluvial fan is unknown. It may be common elsewhere since 'sandy clay' is often recorded in drillers logs, although mud rotary drilling generally prevents identification of such a layer. An indurated sand was subsequently encountered at a site about 20 km to the south, by J. Bell (pers. com., 22.1.00), as part of on-going UNSW-CIC-DLWC investigations.

> *"During coring of the South Tubbo site...a hard sand layer was encountered at approximately 15 m below ground surface...[which] was considerably harder than the matrix above and below.... When the 'cemented sand' layer was encountered the drilling rig almost stalled due to the sudden increase in head torque..."*

The spatial distribution of this clayey sand is a function of its geomorphological origin. Physical, mineralogical and thin section evidence indicated that the clayey sand was hardened by iron-rich clay rather than cemented by either carbonate or silica.

Various modes of origin for iron rich deposits are discussed in detail by Pye (1983), but cannot be differentiated for this deposit with the limited information now available. Iron rich deposits have formed in Australian regolith over a wide range of time through complex landscape evolution (Pain & Ollier 1996), and a number of ferruginised mottled clays sequences in the western Murray Basin of Late Pliestocene and Holocene age are described by Firman (1994). Red, ferruginous sands between 2 and 12 m deep are associated with acidic groundwater in the Shepparton sands of Northern Victoria (Macumber 1991). In that case there were several possible development mechanisms: past lower watertables, current near watertables with more ready access to oxygenated water and past pluvial periods associated with oxygen rich recharge water.

Clay mineralogy of the indurated clayey sand is dominated by kaolinite (Table 5.3) suggesting that accelerated weathering of iron rich minerals may have occurred during a relatively wet climatic period such as the mid-late Pleistocene epoch. At that time, the overlying sand deposits which now form the shallow aquifer would probably have been unsaturated. Since palaeoclimatic conditions would have effected the whole region, the spatial extent of indurated clayey sands within the local area must be determined by the distribution of alluvium containing iron-rich minerals.

5.4 Stratigraphic nomenclature

Some comments about the naming of specific aquitard and aquifer units at the Tubbo site is warranted at this stage. In this thesis, the units are simply referred to as the upper, middle and lower silt units in order to avoid the impression that the particular units described here were representative of the Upper and Lower Shepparton and Calivil Formations at other locations. For the purpose of broad comparision however, it is useful to consider where these units might be assigned.

No lithological distinction between the upper, middle and lower silt units was evident in the geophysical logs of the Tubbo site. The upper silt unit is likely to be equivalent to the Shepparton Formation based upon stratigraphic comparisons with profiles in other parts of the Murrumbidgee. The middle unit may belong to either the Shepparton Formation, if based upon similar mineralogy, or to the Calivil Formation, given identical particle size characteristics. The transition between the formations that has been defined as the top of the middle silt unit (49 m depth) for bore 59113 (Woolley 1991) must be open to question. The lower silt unit is probably a member of the Calivil Formation based upon the white and light grey clay with a high proportion of kaolinite (Brown 1989) that indicates long term weathering under humid conditions in the source area.

The 4 aquifers encountered at the Tubbo site are referred to as the Upper Shepparton (aquifer 1) and Lower Shepparton (aquifers 2 &3) and Calivil (aquifer 4).

5.5 Summary

- Based on mineralogical and litho-stratigraphic evidence obtained from minimally disturbed core, an indurated clay sand (25–32 m) and 3 clayey silt units were identified; a lower unit (75–83 m), a middle unit (45–64 m) and an upper unit (0–16 m).
- Electrical image surveys indicate that the upper unit (at least) is spatially discontinuous, interrupted by large palaeodrainage features probably containing sands and gravels. These palaeodrainage channels are buried beneath a thin veneer of clay and are not apparent at the surface.
- The indurated clayey sand forms a laterally extensive hydraulic barrier at the Tubbo site. This distinctive iron-rich kaolinitic clayey sand, probably also occurs elsewhere reported only as 'clayey sand' in drillers logs, however its regional extent depends on the distribution of alluvium containing iron-rich minerals.
- Despite low ($<10^{-8}$ m/s) permeability of the silty units, leaching of soluble salts from the deep units indicates that some leakage has occurred through the aquitard matrix. Permeability of the silty units was dependent on the scale of testing, suggesting that fracture flow may occur.

Chapter 6

GROUNDWATER HYDRODYNAMICS

Here, groundwater response to pumping from the confined aquifer is assessed in a regional context, and during an irrigation season at the Tubbo site. Hydraulic evidence for flux through aquitard windows is presented, and quantified with a radial axisymmetric flow model.

6.1 Results

6.1.1 Regional groundwater hydrographs

Groundwater levels have been monitored in the area since the 1970s when the first observation bores were installed in confined aquifers. Although groundwater level data that predates irrigation development in the area is not available, a good dataset is available to assess stresses caused by pumping over the medium term. Seasonal trends, however are not always well represented given that monitoring frequency varies between twice a year, and monthly data which is read off chart-recorders.

A detailed region-wide assessment of hydrographic data can be found in Lawson & Webb (1998). Hydrographs for those observation bore sites which were included in the 'snap-shot' chemical sampling are presented in Figure 6.1. All standing water levels (SWL), which by DLWC convention are measured to the top of steel casing, have been converted to relative elevations of the Australian Height Datum (AHD). This permits comparison of flow gradients and directions.

Drawdown due to pumping stresses is seen to increase over time, starting first in the area south of Darlington Point and later at the sites along the river and finally at the CIA pump site. Greatest seasonal drawdown of 15–20 m occurs at the DP south site,

A

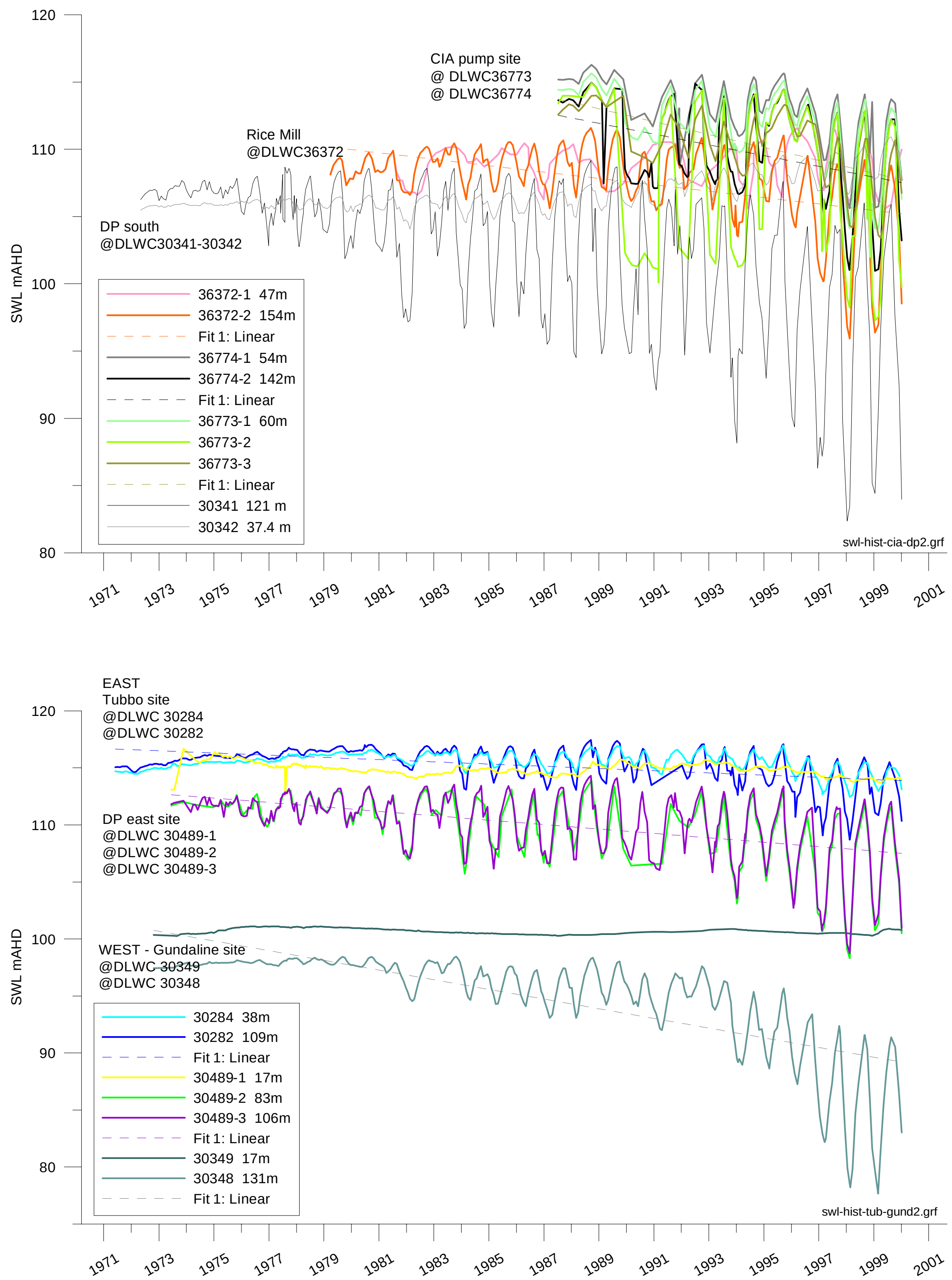


Figure 6.1: Hydrographs for selected nested piezometer sites, 1971–2000 **A** CIA to Darlingtonpoint section **B** Tubbo to Gundaline section.(DLWC data, P. Kumar, pers.com.)

compared with the least marked seasonal drawdowns of less than 5 m at the Tubbo site. Long term drawdown, without recovery at the end of the pumping season is evident from the early 1980s at most sites, but becomes even more marked at the end of the 1990s.

Hydraulic gradients indicate the direction of groundwater flow is towards the west along the Tubbo to Gundaline section and from south to north towards the river from the CIA pump site. The latter gradient is probably a recent development due to rising groundwater levels beneath the CIA prior to 1987 (not shown), resulting in a groundwater mound. The characteristics of this groundwater mound are described elsewhere (eg. Lawson & Webb 1998, Khan et al. 2000).

The direction of flow between the shallow and deep aquifers may also be deduced The lack of trend in 30349 at Gundaline (together with anomalous chemical data that will be presented) pointed to the fact that either the screen of the piezometer was blocked, or that it sampled an isolated pocket aquifer. At the other sites, small upwards gradients were indicated by the hydrographs prior to seasonal drawdown which in most cases has reversed the gradient so that potential flow leakage is downwards. Flow dynamics at the CIA pump site however, appear to be more complex and is under further investigation.

The degree of hydraulic connection between shallow and deep aquifers has often been estimated on the similarity of seasonal drawdown. On this basis, there appears to be good connection betweeen the 38 m and 109 m deep aquifers at Tubbo. At the DP east site, only the deeper two aquifers are connected, while the shallow 17 m aquifer is a very muted reflection of seasonal drawdown. At the other sites, hydrographs for shallow and deep aquifers are more similar, which could be interpreted as hydraulic connection. However, caution is needed in these interpretations for those piezometers screened in clayey material, due to compressible storage effects.

6.1.2 Pumping history at Tubbo

The majority of irrigation bores are location south-west of Darlington Point, many of which are licensed to pump up to 6 500 ML per year. As was noted in Chapter 2.6, there is a significant difference in the volumetric allocation of groundwater and the actual usage. Extensive groundwater pumping has not developed to a large extent in the CIA presumably due to accessible surface water supplies. The CIA pump bore currently holds the largest volumetric allocation within the irrigation area. One of the large volume irrigation bores is the Tubbo bore GW59113 which was installed in 1984. The bore water is stored temporarily in turkey's nest dams, and then used to flood irrigate pasture for cattle and sheep production, with the possibility of conversion to rice production in the

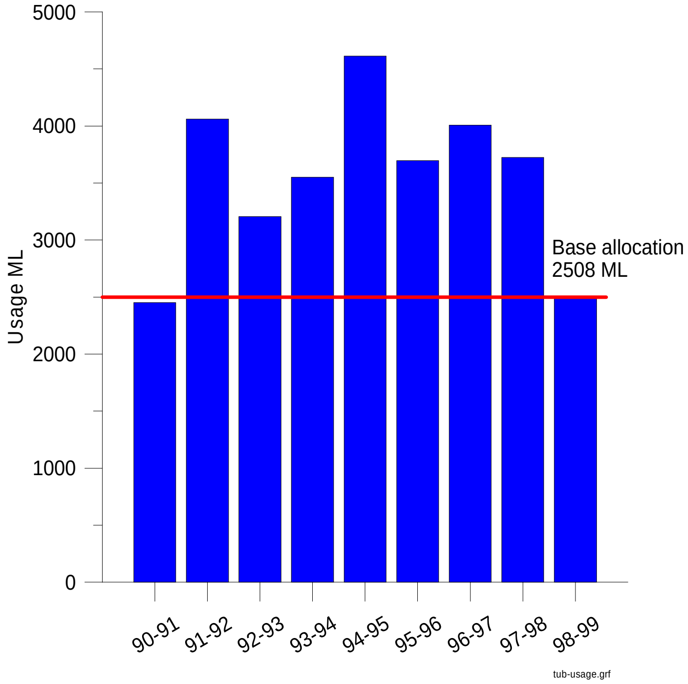


Figure 6.2: Groundwater usage by the Tubbo irrigation bore GW59113, 1990–1998.

future. During the intensive monitoring period at Tubbo, a total volume of ~2 500 ML was pumped from GW59113 which operated intermittently from Oct-98 to May-99 (see pumping records including on CD-ROM). This was less than pumped in previous years (Figure 6.2). Pumping records (J.Gaston, pers. com.) reveal that the greatest volume was extracted during January 1999. Water was applied to irrigated pasture paddocks on a rotating basis.

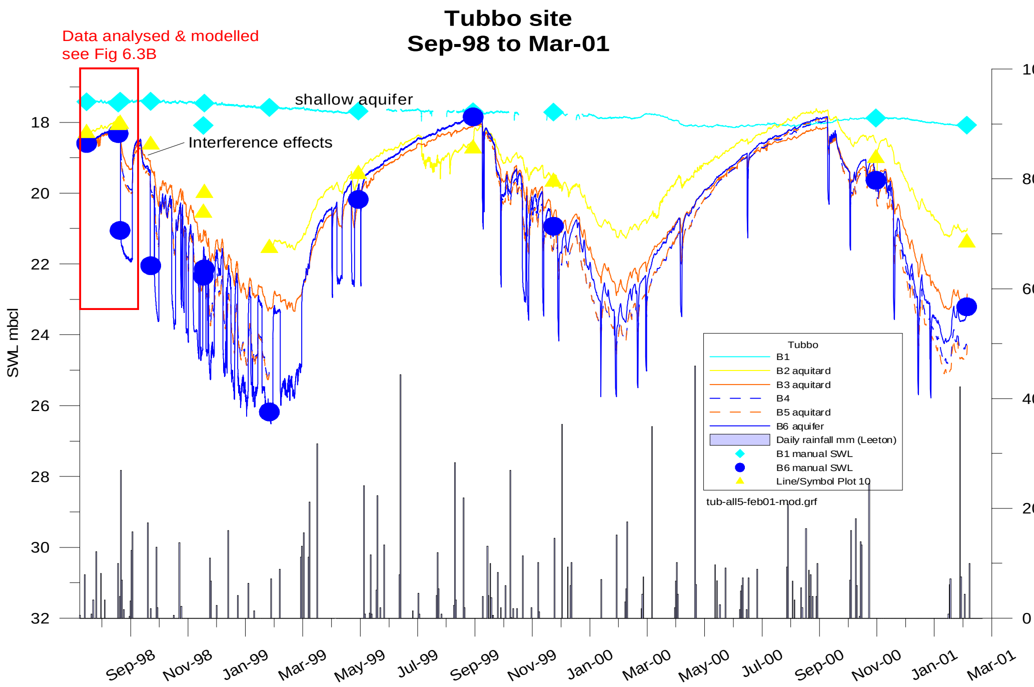
6.1.3 Groundwater levels

Much insight about groundwater flow can be gained from basic interpretation of water level variability. Medium term groundwater fluctuations are now compared with intensive groundwater level data (Figure 6.3) collected from specific depths in the multi-aquifer system between 1998 and 2001.

Two and a half drawdown-recovery cycles are evident in Figure 6.3A. Drawdown in the deep Calivil aquifer is greatest during the first of these cycles, the 1998–1999 season, due to frequent intermittent pumping of the irrigation bore 40 m away from the nested piezometer site.

The increase of water levels in DLWC 30284 between t = 20–500 mins is an unexpected response to pumping (Figure 6.4). This reverse-drawdown response could be

A



B

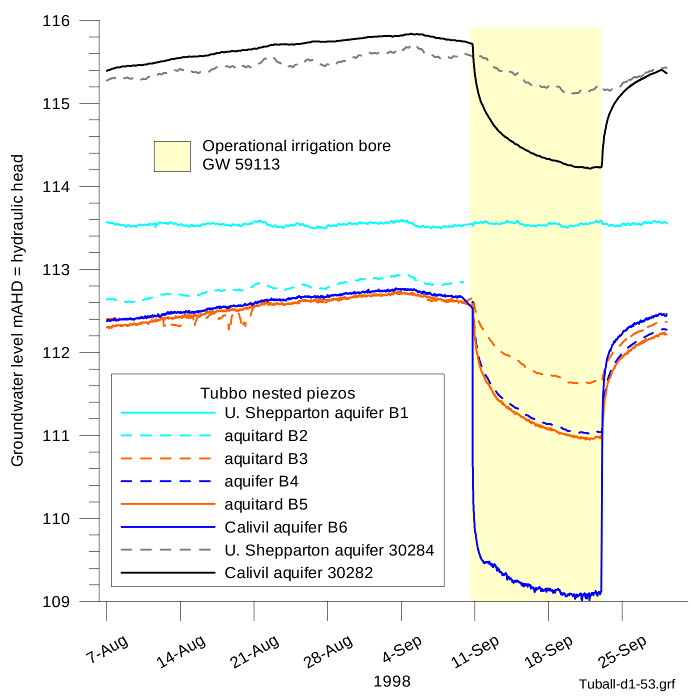


Figure 6.3: Hydrographs for the Tubbo site **A** Sep-98 to Feb-01 **B** Enlargement showing detail for the 1st pumping cycle Sep-98.

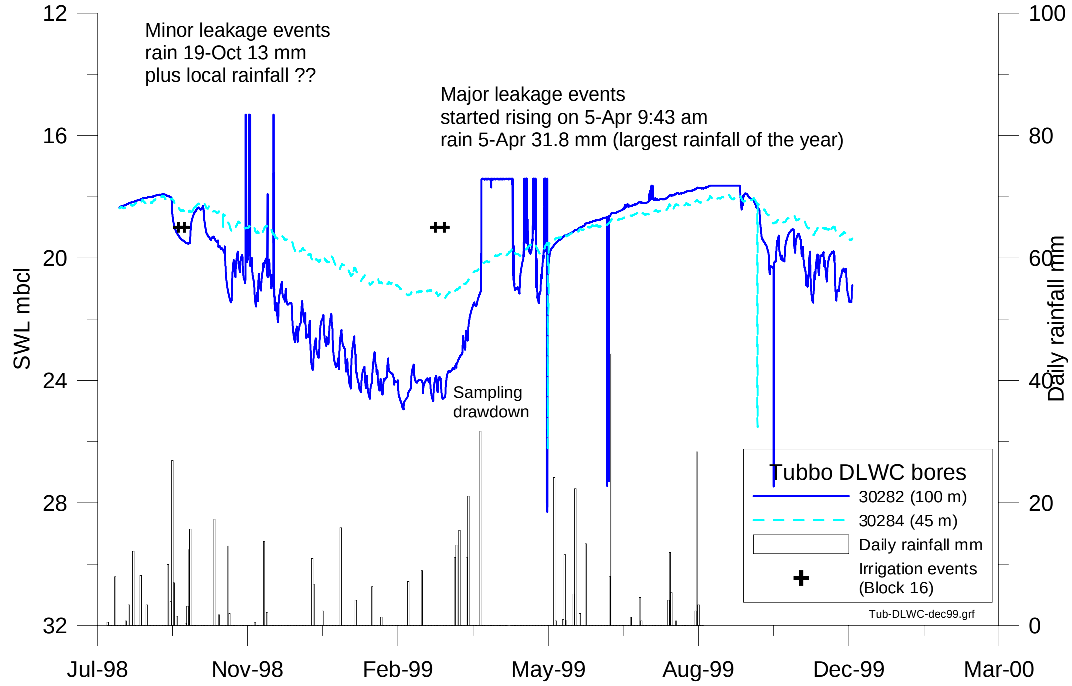


Figure 6.4: Hydrographs for the DLWC monitoring bores 30282, 30284, Tubbo site.

attributed to recharge by irrigation water applied in this paddock. However, irrigation schedules (J. Gaston, pers. com.) showed that this area of the property was not irrigated until 5 days after pumping began. A possible explanation for this phenomena may be the Noordbergum effect which has been documented for layered aquifer systems by **?**). Early stage reverse-drawdown in aquifers and aquitards overlying a pumped aquifer, was attributed to the faster propagation of mechanical (deformation) versus hydraulic (drawdown) stresses due to pumping. This occurs in material that has a high specific storage (compressibility) relative to hydraulic conductivity.

Interference effects

Interference effects on observed groundwater levels are probable due to several nearby irrigation bores (Table 6.1). Ideally, to assess induced leakage around a specific bore, these extraneous influences should be removed, or eliminated by selecting a period of time when only one pump was operational. In a confined aquifer such as the Calivil, drawdown cones will be extensive and will likely coalesce.

Correction for these interference effects is not possible without a lengthy period of water level data over which to determine the effect of each individual bore on regional

Table 6.1: Location of bores, Murrumbidgee River and main irrigation channel, relative to Tubbo irrigation bore GW59113.

Site	Easting	Northing	Distance (km)
Teehan - GW400029	429875	6163280	2.8
Belvedeer - GW65947	431832	6161180	4.5
Caringal - GW59202	434220	6158745	6.8
Murrumbidgee River - closest point			~6.0
Murrumbidgee River - flow line	Euroley	Bridge	~7.5

water levels. This is not possible, since waterlevels in B6 were never at equilibrium. There is however, a short period at the beginning of the 1998–1999 irrigation season when only GW59113 was pumping with a total drawdown of 3.66 m. Later, when only a 2nd irrigation bore (possibly GW400029) was pumping, the drawdown in B6 increased by 0.77 m, or approximately 21% of the drawdown caused by pumping GW59113 alone.

Over subsequent irrigation seasons, when the Tubbo bore was rarely operational, drawdown in the Calivil aquifer was significant (Figure 6.3). In 1999–2000 and also in 2000–2001, drawdown observed in B6 was 23 m, which is only slightly less than the 26 m drawdown observed in the 1998–1999 season.

6.1.4 Gradients

Head differences and potential flow directions, between various layers in the system, based on groundwater levels measured by hand dipper, are shown schematically in Figure 6.5. Gradients, which drive potential flow, are given at key points, assuming head loss over the thickness of each aquitard. Viewing gradients in this manner serves to highlight the dynamic nature of the system in time and space.

Large gradients develop soon after pumping begins in Sep-98 between the Calivil aquitard and aquifer, and much later (Feb-99) between the Upper Shepparton aquifer and clayey sand aquitard. Upwards flow is evident at some points. Vertical gradients were greatest over the indurated clayey sand which separates the Upper and Lower Shepparton aquifers (Figure 6.5). Here, the vertical gradient was 0.1 to 0.67 during periods of quasi-equilbrium and maximum pumping stress respectively. This maximum gradient is similar to a maximum of 0.7 observed across a lacustrine clay aquitard in a multi-layered system below Mexico city (Rudolph et al. 1991).

Horizontal gradients in the Calivil aquifer were at least an order of magnitude smaller than vertical gradients. The horizontal gradient between DLWC-30282 and UNSW-B6 was

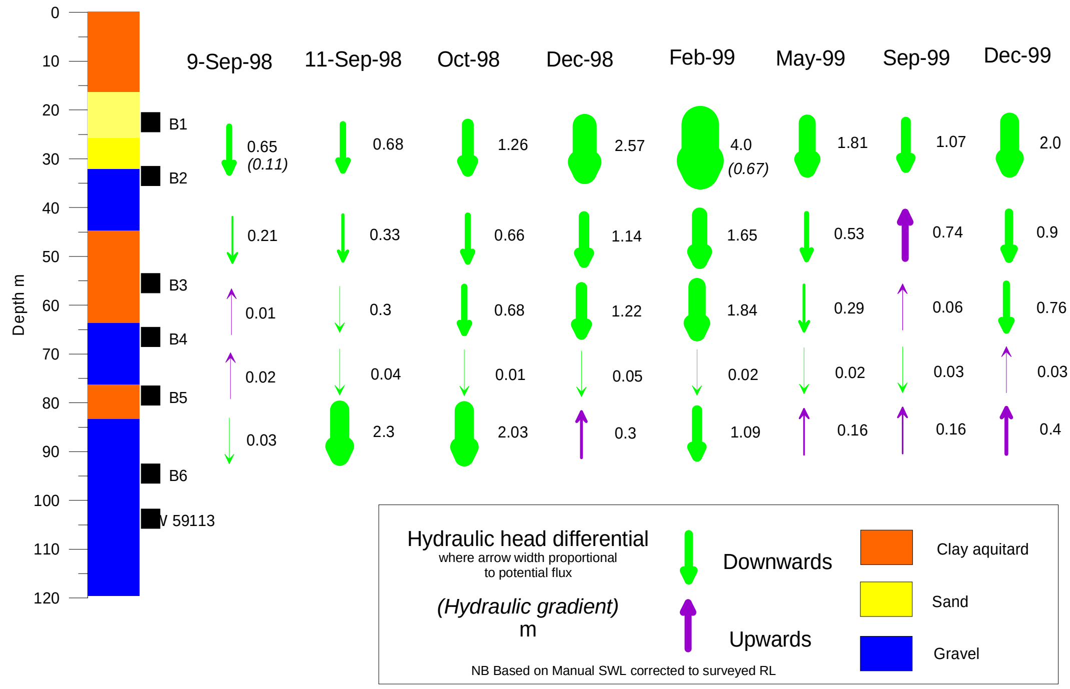


Figure 6.5: Head differences and gradients during the 1998–1999 irrigation season, Tubbo site. See Figure 6.3 for pumping history.

between 0.002 and 0.005 during periods of quasi-equilbrium and maximum pumping stress respectively. These values are similar to horizontal gradients in the semi-arid San Joaquin Valley, reflecting low rates of rcharge to the system (Belitz & Phillips 1995).

As a result of pumping stress, vertical gradient increased by a factor >6, and horizontal gradients by a factor >2. The development of large gradients in response to stress is in itself evidence of low permeability material that has a high resistance to flow. However, increased gradients also serve to increase the rate of flow through the material.

6.1.5 Drawdown analysis

The log-log (dimensionless drawdown) and semi-log drawdown vs time plots (Figure 6.6) for B6 are diagnostic of a confined leaky aquifer (Kruseman & de Ridder 1990). The semi-log drawdown plot is comprised of 4–5 linear steps, each with a decreasing slope. There is an indication of an equilibrium established, with drawdown stabilising after 10000 minutes since the start of pumping. This feature is more distinct on the log-log drawdown plot. However the interpretated equilibrium is based on a limited number of measurements, and must be regarded as tentative since, unfortunately, pasture irrigation had been completed by this stage and the pump was turned off.

Table 6.2: Drawdown delays observed at Tubbo site after start of pumping, 11th September, 1998.

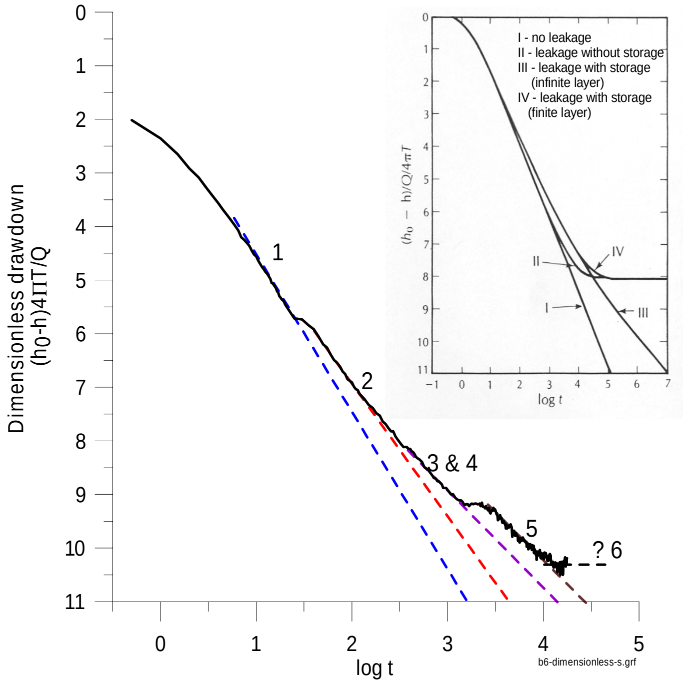
Bore	Drawdown delay after start of pumping (minutes)	Rank
B1	3000 (2.1 days)	7th
B3	57.6	5th
B4	28.8	3rd
B5	43.2	4th
B6	0	1st
30284	∼400 ?	6th
30282	4.4	2nd

Probable sources of groundwater for each section, 1 to 6 on the drawdown curves (Figure 6.6), are now described in the context of drawdown times observed in overlying aquifers (Table 6.2).

- 1. (t = 0 mins) Water sourced from Calivil aquifer
- 2. (t = 25 mins) Additional water sourced from L. Shep/Calivil aquifer, and soon after (t = 29 mins) drawdown is observed at B4.
- 3. (t = ∼150 mins) Additional water sourced from Lower Shepparton aquifer.
- 4. (t= 500 mins) Additional water sourced from distant Lower and/or Upper Shepparton aquifer (at DWLC 30284).
- 5. (t= 1 800 mins) Additional water sourced from close Upper Shepperton aquifer, and soon after (t = 3000) drawdown is observed at B1.
- 6. (t = + 10 000 mins) Possible quasi-steady state, due to recharge boundary, though insufficient evidence for definitive identification.

A further note regarding features of interest on the drawdown plots. At some steps (1–2, 4–5), drawdown ceases for a short time as additional water is sourced, and is thereafter re-established, but at a lower rate. Drawdown in B1 of about 2–3 cm that was observed at the same time as B5, which was installed in the same drill hole, is thought to be an artifact of pressure propagation within the installation. Sustained drawdown in B1, that is thought to reflect changes within the Upper Shepparton aquifer at this point, did not occur until t = 3 000 seconds.

A



B

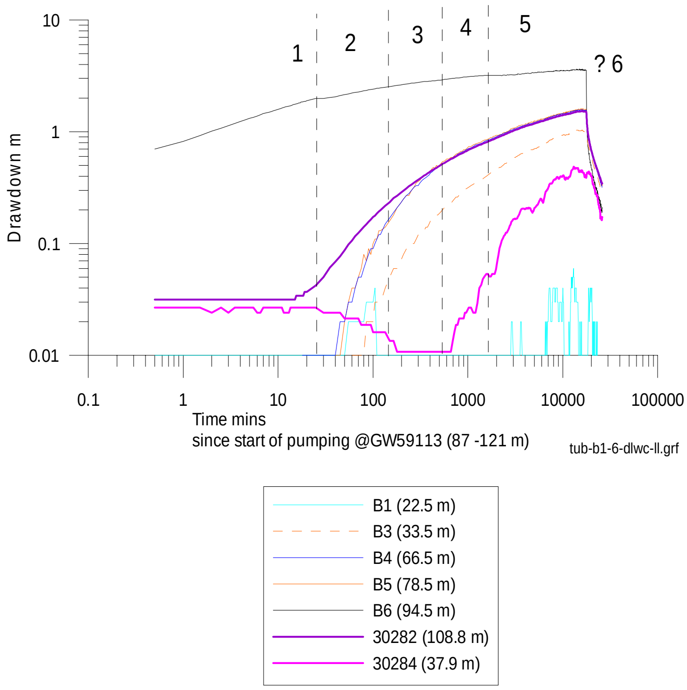


Figure 6.6: Drawdown during pump testing at Tubbo site, 11–24 September 1998. **A** Semi-log dimensionless drawdown showing a series of straightline drawdowns indicating leakage without aquitard storage and steps due to window leakage INSET: Dimensionless drawdown for various leakage scenarios (Source: Fetter 1994 after Hantush 1960). **B** Log-log drawdown.

6.1.6 Pump test analysis

Assumptions

All analytical pump test interpretation methods assume a layer cake stratigraphy, in that layers are spatially continuous and infinite in extent, and that leakage is proportional to change in pressure head (Kruseman & de Ridder 1990). This is a serious limitation in a fluvial fan environment where aquitards are known to be spatially discontinuous.

Also, multi-layered systems can not be resolved by simple analytical techniques. The two techniques for interpreting such systems that are offered by Kruseman & de Ridder (1990) require a computer, and are limited to applications where fully penetrating bores pump from more than one aquifer.

Bore storage effects do not appear to be important in this case since drawdown is linear after 1-2 minutes. This was not expected to be a significant factor given the small diameter (50 mm) piezometers. Effects of variations of pumping rate are unknown and assumed to be negligible, particularly given the very large pumping rate (280 L/s) of the turbine pump which was driven by a CAT diesel engine. Partial penetration effects do not appear to be significant, as is expected since the screened interval of the pumping bore GW59113 extends over most of the Calivil aquifer.

Aquitard storage was assumed to be negligible in both of the analysis methods selected. If aquitard storage were important, the resultant K for the Calivil aquifer would be overestimated, though the near linear drawdown evident in Figure 6.6 mean that this is probably not a major factor.

Hydraulic conductivity of the Calivil aquifer

Two straight line methods were applied to determine the K_h of the confined Calivil aquifer at B6. Straight line methods were used in preference to curve matching, because of the ease of graphical interpretation that minimises potential error. The methods selected were Jacobs straight line method, and Hantush inflection-point method as outlined by Kruseman & de Ridder (1990), both of which were also applied by Lawson & van der Lely (1992) in analysis of the CIA pump test.

Both methods of analysis were restricted to early time data for section 1, prior to leakage effects from overlying layers. The analysis was performed relative to the pumping bore GW59113 with a pumping rate, Q, of 24 000 m^3/d (280 L/s). It was assumed that the thickness of the Calivil aquifer was 35 m.

In applying the Hantush inflection-point method (for details see analysis included on the CD-ROM), there was some uncertainty in the projected final drawdown, s_m. Nevertheless, the resulting line of inflection was centered on the straight line segment 1 which was a suitable time point for the analysis. The resulting K_h was 114 m/day.

Jacobs straight line method (1946), for unsteady flow in confined aquifer is based on the Theis equation Kruseman & de Ridder (1990). It assumes that u <0.1 (where u = r2/4KDt), which means that the radius must be small and time large for any application to ensure a straight line drawdown relationship. By this method, using constant r, KD was calculated to be 5 490 m^2/day, which for a 35 m thick aquifer is equivalent to a K_h of 156 m/day. Storativity was found to be 1.4×10^{-3}. Back calculations confirmed that u <0.1.

K_h is significantly higher than that determined in the CIA (Lawson & van der Lely 1992) of between 42–62 m/day. The higher K_h at the Tubbo site may be related to more significant palaeochannel development in the Tubbo area or the closer proximity of the site to the apex of the Lower Murrumbidgee alluvial fan at Narrandera.

Several studies of K in fluvial fan environments have reported systematic hydraulic conductivity variation in proximal and distal fan facies. McCloskey & Finnemore (1996) reported that lower K were expected in the proximal area due to poorer sorting. Poorer sorting is indeed evident in the mixed sand-gravels at DLWC-30282 compared with sand at UNSW-B6 for the Calivil aquifer. The variation of K may be within the analytical error of interpretation.

6.1.7 Slug tests of aquifers

Compressed air slug tests enabled the hydraulic conductivity of the shallow aquifer to be determined, but were of limited use in determining aquifer permeability at depth.

Oscillating recovery curves were observed for B1, B4 and to a minor extent in B6 (see hydrographs on the accompanying CD-ROM). These were attributed to inertial effects that characterise highly permeable aquifers and deep wells (Kruseman & de Ridder 1990). Available methods for interpreting oscillation tests (eg. Uffink) were not suitable for partially penetrating piezometers such as these. However, the Hvorslev method was used to approximate the hydraulic conductivity for the shallow aquifer (B1) of 4.7×10^{-4} m/s (~41 m/day).

6.2 Evidence for window leakage

Hydraulic evidence indicate that leakage between the Lower Shepparton and Calivil aquifers occurs through aquitard windows.

The clearest evidence was that drawdown was observed first in the aquifer above the Calivil, prior drawdown occurring in the intervening Calivil aquitard, suggesting that the pressure change had propagated around, rather than through the aquitard matrix.

Secondly, if leakage occurred through the aquitard matrix, drawdown would be observed in the overlying aquifers at approximately the same time as the drawdown step as additional water was sourced. This is because, aquitard leakage (or fracture flow) would be expected greatest close to the pumping bore which induced downwards gradients. However, a time delay is evident between these steps and drawdown observed at B4 in the Lower Shepparton-Calivil aquifer, and at B1 in the Upper Shepparton aquifer. The time delay is attributed to the time required for lateral propagation of pressure change within each aquifer.

Finally, head differences in Figure 6.5 indicate both upwards and downwards gradients in Sep-99. At this time, there was potential for upwards flow between each of the lower aquifers, although a downwards gradient is indicated between the Calivil aquitard itself and the Calivil aquifer. Pressure levels observed in B5 were clearly not in equilibrium with those in the aquifers.

Evidence for leakage through aquitard windows is consistent with the geomorphological origin of these deposits, as discussed in Chapter 5. Fluvial deposits are characterised by discontinuous interchannel deposits, particularly on an alluvial fan. By contrast, aquitard deposits from lacustrine environments would be laterally extensive. Hydrochemical evidence for rapid downwards migration of solutes at depth during pumping, is also to be presented.

6.3 Profile and radial flow modelling

Quantifying vertical leakage due to pumping in a complex multi-layered system is best achieved by a numerical flow model that is constrained by independent data. Numerical modelling is the only feasible method of realistically examining leakage through aquitard windows, that have been shown to exist at Tubbo site by hydraulic and hydrochemical data.

Profile and axisymmetric modeling is introduced by Anderson & Woessner (1992). Profile modelling is best suited to situations where vertical flow is important but a full 3D

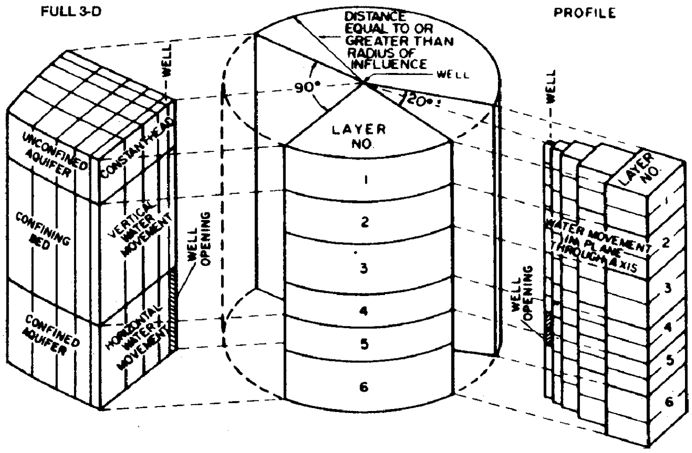


Figure 6.7: Radial axisymmetric flow modelling (Land 1977).

model is not possible because of a lack of data or project constraints. Other applications include testing the validity of conceptual model, prior to full 3D modelling, and to study regional cross-sectional flow. All flow is assumed to be parallel to and in the plane of the profile, and must be aligned along the flow line. Where flow occurs at an angle to the profile (eg. towards a well), axisymmetric profiles must be used (Figure 6.7).

6.3.1 Model purpose

The purpose of numerical modelling was interpretative, rather than predictive, with the goal to quantify vertical leakage versus lateral flow to a pumping bore. Since it is common for flow models to be non-unique (Barrash & Dougherty 1997), independent evidence should contribute to model development (Bath & Lalieux 1997). The limitation of numerical models was discussed by Ghassemi et al. (2000) who stated that:

> *" our ability to develop computer codes capable of simulating complex systems now exceeds our ability to supply the input data necessary for reliable calibration."*

The absence of permeability data is a major reason why model outcomes are non-unique (Neuman et al. 1980). In numerical models, aquitards are often not explicitly included. Rather, inter-aquifer leakage is based upon a lumped parameter, leakance, ℓ. As shown in the equations below, leakance is determined from aquitard thickness (from drilling records), and estimated aquitard K_v. However, leakance is most sensitive to aquitard K_v which varies over several orders of magnitude, and inadequate hydraulic parameter estimation may not become evident until stresses are impossed on the system.

$$\ell = \frac{K'_v}{D'} = \frac{1}{c} \tag{6.1}$$

$$L = \sqrt{KDc} \tag{6.2}$$

Where ℓ is leakance (1/day), L is leakage factor (m), K'$_v$ is vertical hydraulic conductivity of aquitard (m/day), D' is thickness of aquitard (m) and c is hydraulic resistance (day).

The Tubbo site flow model explicity includes aquitard units. The flow model benefited from measured hydraulic parameters (K_v and S_s) for the aquitards, and K_h for the Upper Shepparton and Calivil aquifers determined by slug test and pump test analysis of early drawdown data respectively.

Quantifying vertical leakage versus lateral flow, was attempted then, by developing a numerical flow model, based on a realistic conceptual model, and constrained with measured hydraulic parameters and groundwater levels. Independent verification of the model using hydrochemical and isotopic data is presented in Chapter 8.4.

6.3.2 Conceptual model

Of the three conceptual models proposed for an aquifer-aquitard system in Figure 1.1, field investigations demonstrated that aquitard windows should be included in a realistic flow model for the Tubbo site. By contrast, a simple layered conceptual model would be adopted for a regional scale flow model, given that whole Tubbo site would lie within only one or two 1.25 km^2 cells.

Numerical flow models were developed using the generalised setup depicted in Figure 6.8 that adequately simulated hydraulic heads observed in the Tubbo nested piezometers (Figure 6.9). The conceptual model included aquitard windows through layers 4 and 5 and a laterally continuous aquitard (layer 2 - the indurated clayey sand) to separate the Upper and Lower Shepparton aquifers. Although aquitard windows have been shown to exist near the surface through the vadose zone (not modelled), and in the middle (layer 4) and lower (layer 5) silt units, the width and hydraulic conductivity of the windows, and location relative to the pumping bore was unknown. A number of numerical models were developed that assessed the impact of these variables, including leakage around the bore casing itself, relative to leakage in a simple layered system without aquitard windows.

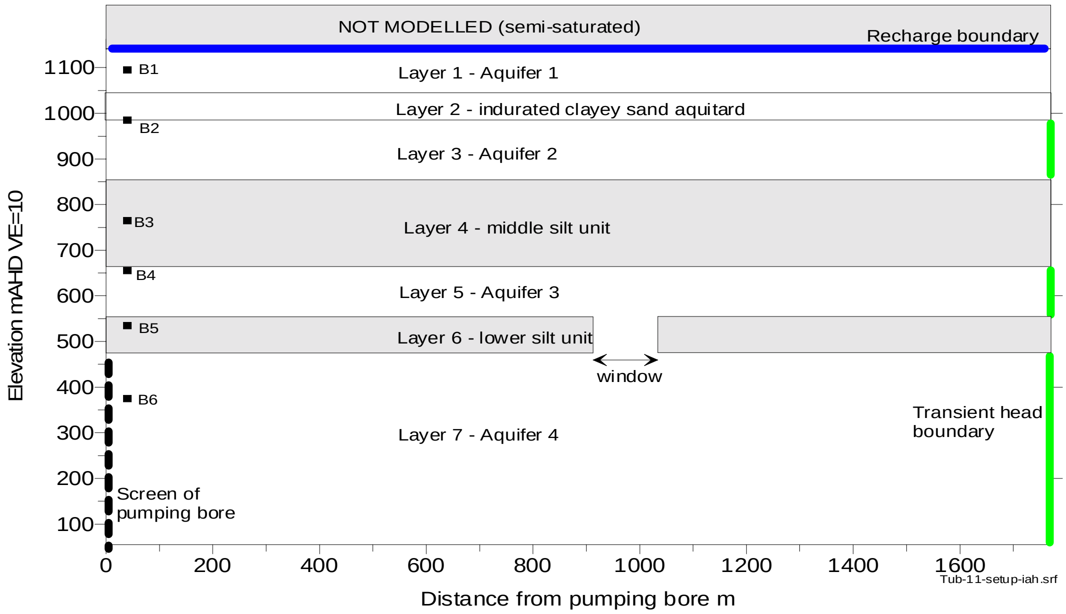


Figure 6.8: Setup of layers and boundary conditions for FEFLOW modelling at Tubbo site.

6.3.3 Model development

Numerical flow models were developed using the finite element subsurface FEFLOW (Finite Element subsurface FLOW) system developed by the German WASY Institute (Diersch 1998). FEFLOW is a user-friendly modeling system for 2D, 3D or cross-sectional modelling of density-coupled or uncoupled, variably saturated, transient flow and chemical transport in groundwater. Finite-element methods are used to describe groundwater flow dynamics, including free surfaces (perched water table), and convective and dispersive contaminant transport. It is capable of modelling mixed conditions due to surface water interactions or pumping, confined and unconfined multi-level aquifers and transient boundary conditions in regard to flow or leaching sources.

Extensive example problems are supplied in the FEFLOW manual that verify approximations such as spatial discretiztion and iteration schemes and the iterative solver (Diersch 1998). Cross-verification with other models was reported by Kolditz et al. (1998) for density flow problems including upconing of saline brine.

Steady state profile model

A series of models of increasing complexity were constructed, beginning with 2–3 layer steady-state profile models in order to test the sensitivity of various parameters. A 7 layer profile model was then constructed with the same dimensions and layer thicknesses of the

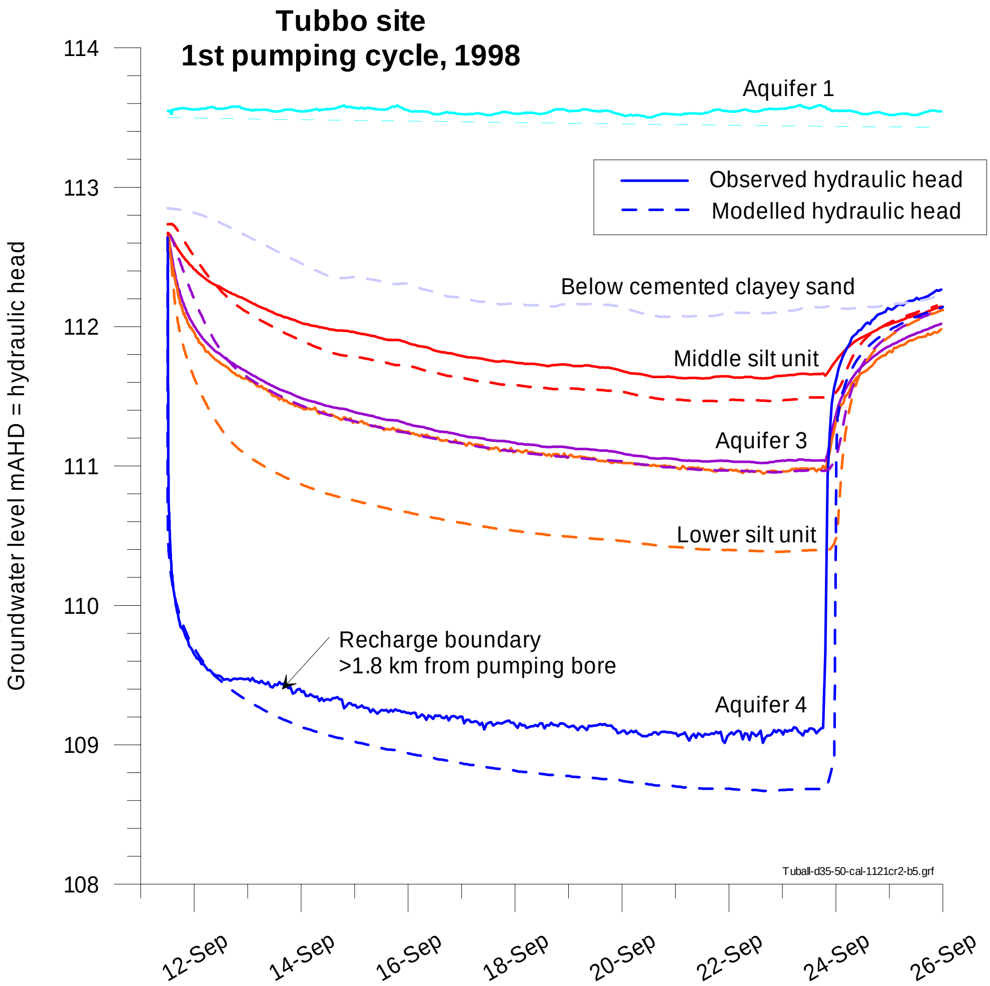


Figure 6.9: Observed and modelled hydraulic heads at Tubbo during the 1st pumping cycle, 11–26th September, 1998. (Model: 11.21cr2).

intended axisymmetric model, in order to generate initial hydraulic heads (for Sep-98) throughout the model domain for transient modelling. For this purpose, the aquitard layers were assumed to be laterally extensive. Vertical groundwater heads at the Tubbo nested piezometers (Figure 5.8) were extended into 2D with data from the DLWC bores, 1.77 km upgradient.

Multiple solutions could be generated that achieved the observed head differences. These specified relative aquitard permeability for layers 2, 3 and 5 (eg. K 10^{-11}, 10^{-10}, 10^{-10} m/s respectively). Values of K between 10^{-11} and 10^{-6} for a specific layer resulted in the same vertical head distribution, but the magnitude of vertical leakage through these layers clearly varied greatly as a result. This finding reinforced the need to use meausured permeability data for the aquitard layers.

Transient axisymmeteric model

Transient axisymmetric models were then prepared, centered on the irrigation bore GW59113. The final axisymetric transient flow model comprised of 4 aquifers and 3 aquitard units (7 layers) over the 110 m thick sequence at Tubbo, using a vertical exageration of 10. A 15 day period (11–26th September, 1998), of the first pumping drawdown and recovery cycle (Figure 6.3) was modelled, for reasons previously discussed. Of this time period, 0.5 day was quasi-equilibrium, 12.5 days pumping and 2 days recovery.

Automatic time step control was implemented by FEFLOW that resulted in steps of <1 sec during the early stages of pumping. The execution time of these models on a personal computer ranged from 0.5 to 1.0 hour, and for the model with variable element width zones, execution time of several hours.

An example of hydraulic response to pumping in the multi-layered system, at one time step using this FEFLOW model setup is shown in Figure 6.10.

Mesh development

Mesh quadrangulation (rectangular elements) was adopted. Most of the models developed were based on a radially expanding element width (10594 elements and 10801 nodes). In RADMOD (Reilly & Harbaugh 1993), the cell expansion factor, was defined as:

$$\alpha = \frac{r_{i+1}}{r_i} \tag{6.3}$$

Where r is the radial distance to a node (m), i is the index number of the column, and α generally varies between 1.2 and 1.5.

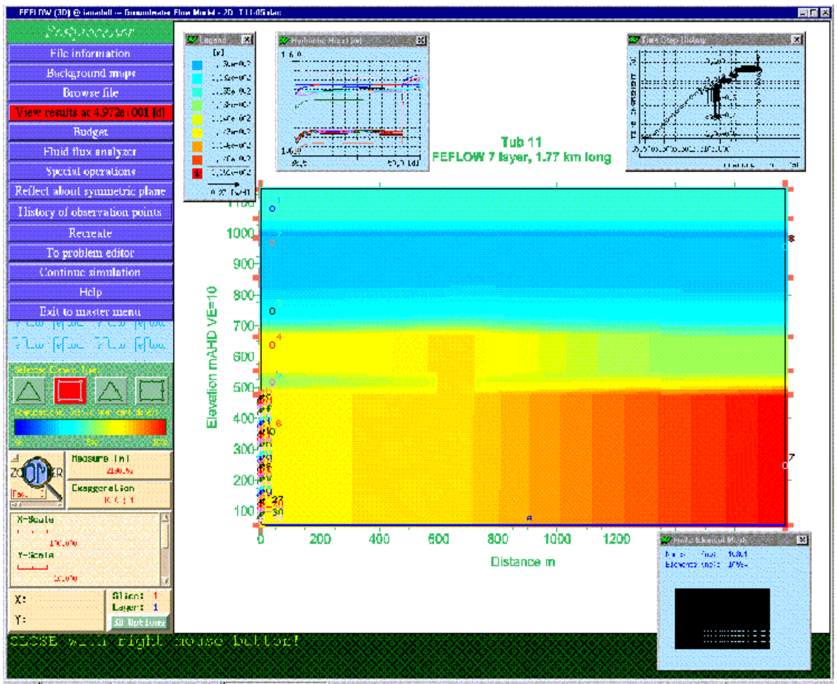


Figure 6.10: Example of axisymmetric FEFLOW model for a multi-layered system with an aquitard window, showing hydraulic response to pumping stress.

However, due to the method of constructing superelements within FEFLOW, the cell increase factor, α was not constant, but decreased with increasing radially distance (α 2.1–1.4). By configuring very small cell widths (<0.1 m) near the pumping bore, numerical discretization errors were kept to a minimum. Mesh enrichment near the pumping bore screen, ensured that sufficient quantities of water were able to be transmitted to the bore without unrealistic drawdown (Figure 6.11).

In order to assess leakage around the bore casing, and very narrow window leakage, an alternative rectangular mesh was constructed with constant cell width in each of 4 zones. This enabled very narrow cells (<0.01 m) near the pumping bore, and either side of the nested piezometers (r=35 m), but with wider cells elsewhere so as to minimise required computing power. Nevertheless, the large model (40 044 elements, 40 469 nodes), required hours to execute.

Comparison of drawdowns generated by identical models, but with the regularly expanding mesh (model Tub-1121c), and variable mesh blocks (model Tub-30me) resulted in minor differences during the first day of pumping. No major differences over the 15 day period were evident, so it was concluded that mesh effects were not significant.

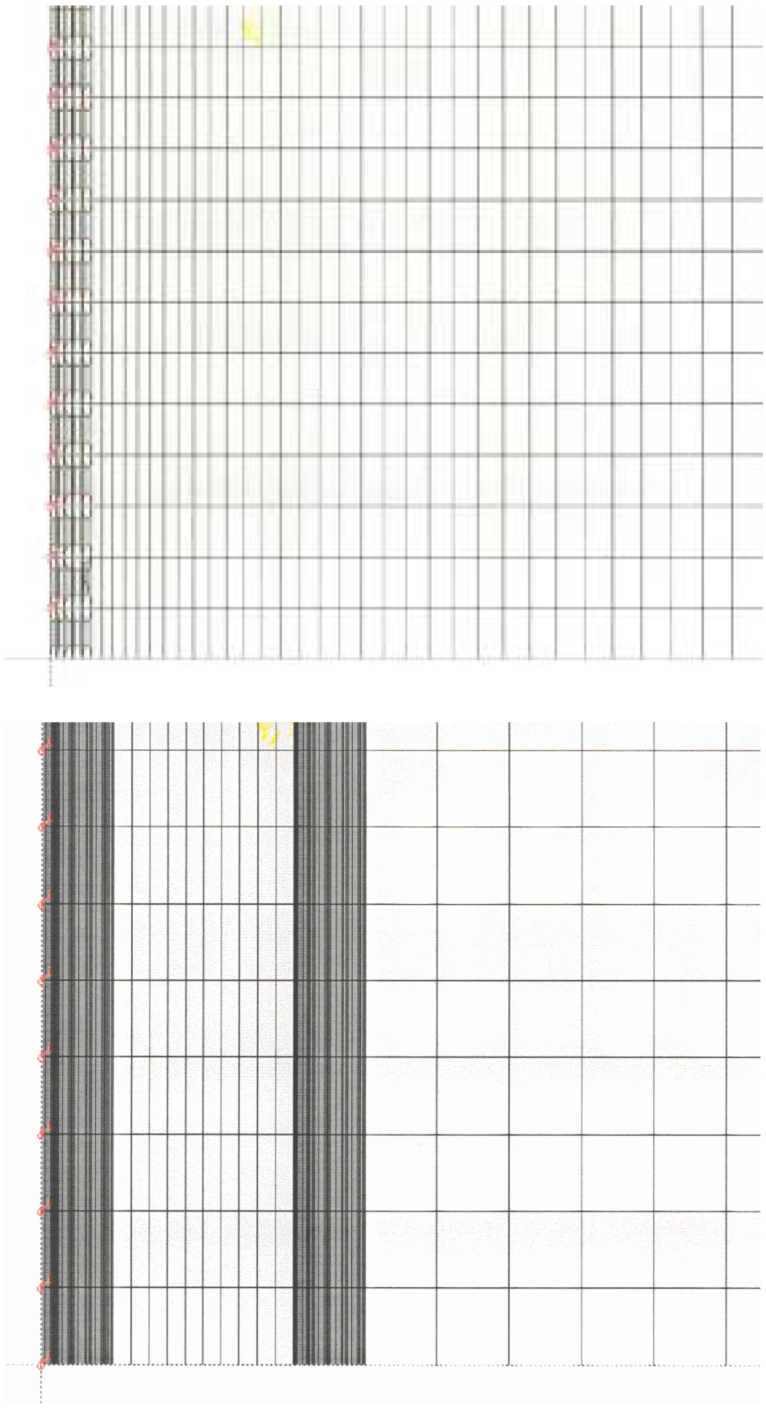


Figure 6.11: Mesh setup showing radially expanding elements and mesh enrichment near the pumping bore (model Tub-1121c).

Boundary conditions

Flow boundaries were of both the Neumann specified flow and Dirichlet specified head types.

Lateral flow into the model was determined by transient head boundaries. These were established at a radial distance of 1.77 km for aquifer layers 3 and 7, using the record from data loggers installed in bores 30282 and 30284 respectively. A transient head boundary for aquifer layer 5 was configured equivalent to 30282, since the initial heads in Sep-98 were very similar.

An outflow boundary was established along the axis of layer 7 where outflow volume at each node, q, to represent flow to the pump well, was determined by the following equation:

$$q = \frac{Q}{2\pi N} \tag{6.4}$$

Where Q is total volume pumped (m^3/day), q is each node, and N is number of nodes.

Each node was described by a time-varying function that effectively turned the pump on and off and a hydraulic head constraint that prevented dewatering of the bore.

No flow was specified through the base of layer 4, which was established at the known base of the Calivil aquifer, since the Renmark aquifer is not known to exist at this location. Upwards flow from bedrock was assumed to be zero.

Recharge through the vadose zone to layer 1 was assumed to be nil for the 15 day period of modelling. Since no rainfall occurred during this period, recharge via rapid flow pathways would not have occurred. Minor matrix flux may have occurred, but was not quantified by this project.

Hydraulic parameters

Hydraulic parameters for aquifer layers were estimated on the basis of previous pumping test and models for the Riverine plains (Tables 6.3, 6.4). Hydraulic conductivity of the Calivil aquifer was set at $\sim$110 m/day, on the basis of pump testing interpretation that has been presented.

Although a range of aquitard hydraulic parameters were initially tested (Table 6.5), the final model was based parameters closest to measured aquitard permeability and specific storage for the aquitards. K values determined by falling head permeameter testing and consolidation testing were used, although the permeability constrast between aquifers

Table 6.3: Hydraulic properties of aquifers in the Murray Basin and Lower Murrumbidgee areas.

K (m/day)	Comments	Reference
Shepparton aquifer		
25–10	sandier zones	Evans & Kellet 1989
0.1–1	cells for Tubbo site	Punthakey et al. 1994
45–50	average	
20–30	Campaspe Valley, Victoria	Brinkley & Reid 1990
50–60	estimated for Darlington Point area	Berhanu 1991
15	estimated away from Murrumbidgee River	
Calivil aquifer		
230	Victoria	Evans & Kellet 1989
2–10	cells for Tubbo site	Punthakey et al. 1994
49–132	Campaspe Valley, Victoria	Brinkley & Reid 1990
24–45	30 day pump test near Elmore, Victoria	
20-28	sieve analysis at Elmore site	
150	estimated at Darlington Point	Berhanu 1991
80–100	estimated west of Darlington Point	
50	468 pump test at CIA	Lawson 1992

Table 6.4: Storativity of aquifers in the Murray Basin and Lower Murrumbidgee areas.

Storativity (Specific storage)	Comment	Reference
Shepparton aquifer		
$1.7\times10^{-4} - 4\times10^{-2}$	ancestral river aquifers	van der Lelij 1978
2×10^{-4}–4×10^{-2}	prior stream aquifers	
2×10^{-4}–8.1×10^{-3}	floodplain aquifers	
2×10^{-2}–2×10^{-1}	Campaspe Valley, Victoria	Brinkley & Reid 1990
1×10^{-2}–2×10^{-1}	representative near Yanco Weir	Berhanu 1991
0.05 - 0.15	Specific Yield	Punthakey et al. 1994
Calivil aquifer		
5×10^{-4}–5×10^{-3}	Campaspe Valley, Victoria	Brinkley & Reid 1990
5×10^{-4}	30 day pumping test at Elmore, Victoria	
1×10^{-5}	respresentative regional value	Berhanu 1991
5×10^{-4}	468 day CIA pump test	Lawson 1992
$(1\times10^{-4} - 1\times10^{-2})$	modelled for Calivil aquifer	Punthakey et al. 1994

Table 6.5: Range of hydraulic properties tested during calibration of Tubbo model.

Layer	K range (m/s)	Ss	Comment
1,3,5,7(aquifers)	2×10^{-11} – 1×10^{-7}	10^{-6}–10^{-3}	
2 (aquitard)	10^{-11}–10^{-07}	10^{-5}	spatially continuous
4 (aquitard)	2×10^{-11}–10^{-07}	10^{-6}–10^{-3}	aquitard windows
6 (aquitard)	10^{-11}–10^{-07}	10^{-6}–10^{-3}	aquitard windows

and aquitards was more important in directing flow through windows than the absolute values of aquitard permeability.

6.3.4 Calibration

Calibration proceeded on a trial-and-error basis. This was very time-consuming but informative in terms of developing an understanding of the complexities of groundwater flow in the system. Groundwater level measurements for piezometers B1-B6, obtained by automated logger, were used to calibrate the model (only the first data point of each was used to generate intial groundwater head distribution). These values were converted to SWL AHD and referenced for the mid-point depth of the piezometer screen.

Calibration with observed heads was best for the aquifers. Aquitard calibration was improved by using the top of gravel packing above the screen as the reference depth, rather than the mid-point of the screen. This difference reflected the rapid loss of head through the aquitard, that contrasted with relatively constant groundwater pressures within an aquifer layer.

A good calibration of the FEFLOW model, as judged by visual comparison, was achieved with observed heads in the middle and deep aquifer as shown in Figure 6.9. However, this could be achieved either by spatially continuous silt units with a K less than 10^{-7} m/s, or silt units with K of 10^{-9} m/s which were spatially discontinuous. This indicates that hydraulic head measurements at specific points within the multi-layered system, may not provide any indication of rapid flow that occurs through localised aquitard windows. A similar finding was reported by Fogg (1986) for a palaeochannel which had a dominant effect on flow, as indicated by velocity vectors, but little influence on hydraulic head distribution of the model.

A more quantitative measure of calibration were determined for each aquifer, and the Calivil aquitard at two points in time, by mean error (ME) and root mean squared error (RMS). These are defined by Anderson & Woessner (1992) as follows:

$$ME = \frac{1}{n}\sum_{i=1}^{n}(h_m - h_s)i \tag{6.5}$$

$$RMS = \left[\frac{1}{n}\sum_{i=1}^{n}(h_m - h_s)i^2\right]^{0.5} \tag{6.6}$$

Where n is the number of calibration values, h_m is measured head and h_s is simulated head.

For the best[1] model, at t = 36.01 days, the mean error was -0.15 and 0.06, with and without the Calivil aquitard respectively. At t = 38.51 days, the mean error was 0.35 and 0.13, with and without the layer 6 aquitard respectively.

At t = 36.01 days, the root mean squared error was 0.4 and 0.12, with and without the layer 6 aquitard respectively. At t = 38.51 days, the root mean squared error was 0.51 and 0.16, with and without the layer 6 aquitard respectively.

While these lumped measures of calibration are favourable, the actual error may be lesser or greater at different points in time, and at varying radial distances from the pumping bore. For example, the stepped drawdown pattern that was observed at early pumping stages was not able to be replicated by the flow model (Figure 6.6). An extended calibration period would be required to more precisely simulate drawdown by trialing the countless combinations of aquitard window diameters and locations relative to the pumping bore.

6.4 Modelled water balance

The water balance of the simulation which best matched modelled heads with observed heads is presented in Figure 6.12. For scenarios without aquitard windows, lateral flow accounted for ∼86% of the volume abstracted by the irrigation bore, compared to ∼12% from vertical leakage through the lower aquitard, and ∼2% through the indurated clayey sand. Some water was also sourced from storage, within the lower aquifer and aquitard.

Leakage proportions for scenarios that included aquitard windows are presented in Figure 6.13. The flux ranges given are for numerous individual models and serve to highlight the possible importance of various vertical leakage pathways. By introducing aquitard windows in layer 4 & 6, lateral flow accounted for a smaller proportion (60–90%) of the volume abstracted, compared with vertical leakage of upto 40%. However, most of the vertical leakage to the deep aquifer, was from the middle aquifer, with 1–15% sourced

[1]Model: Tub-1121c

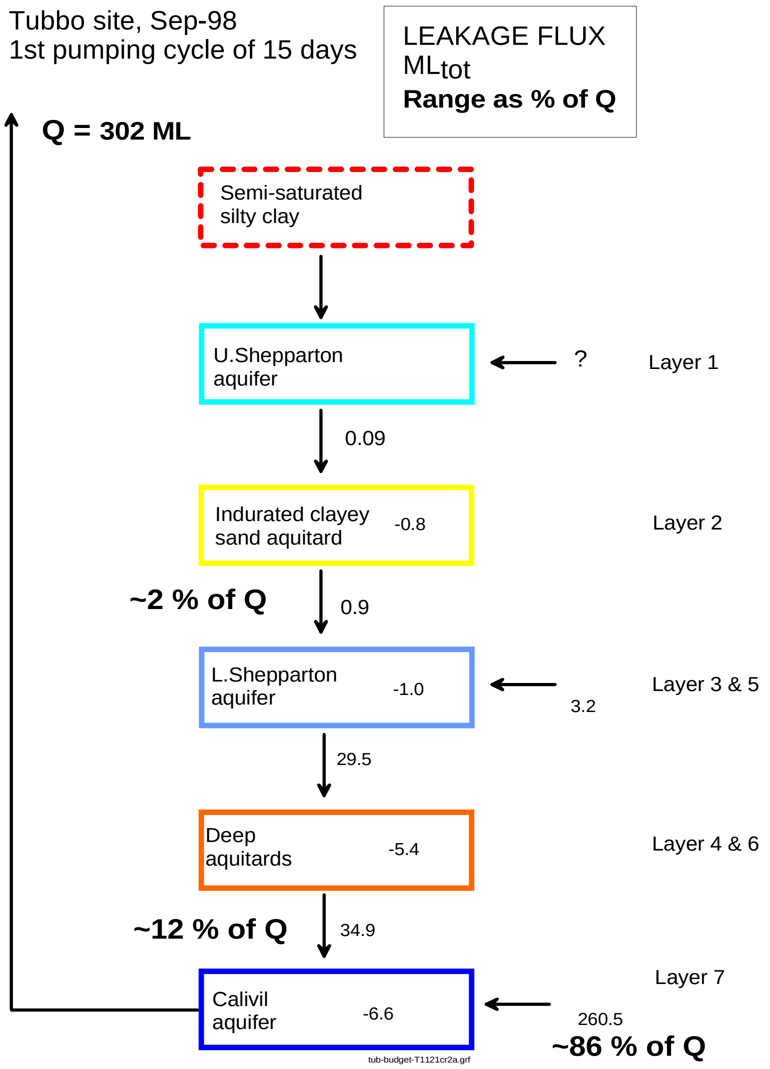


Figure 6.12: Water balance for 1st pumping cycle, with no aquitard windows, 11–26th September, 1998.

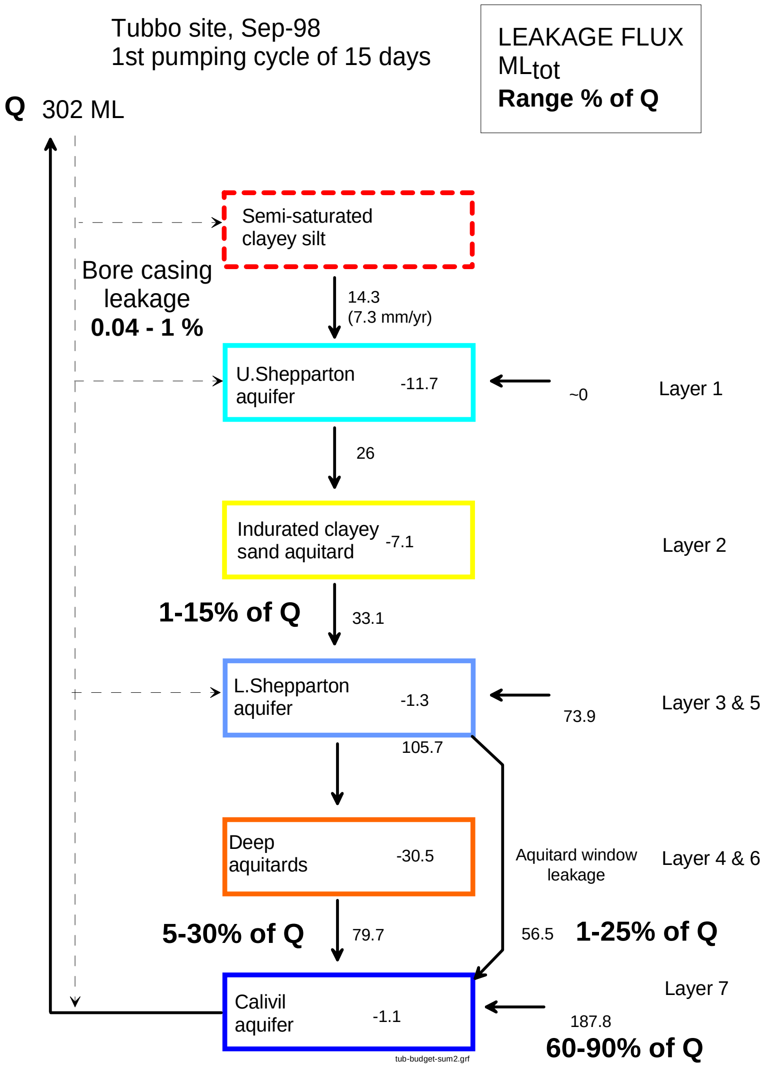


Figure 6.13: Water balance for 1st pumping cycle, with aquitard windows, 11–26th September, 1998. The range of flux specified for each flowpath is a summary of numerous individual models.

from the shallow aquifer. The proportion of vertical leakage that occurred through the lower aquitards was about the same as leakage through aquitard windows, depending on aquitard permeability, window width, location relative to the pumping bore, and other factors. Bore casing leakage accounted for a small proportion of vertical leakage (<1%), relative to leakage through the aquitards and aquitard windows.

6.4.1 Sensitivity to aquitard permeability and compressibility

Sensitivity of modelled drawdown to aquitard permeability and compressibility is depicted by Figures 6.14 and 6.15).

The large drawdown observed in piezometers that were screened in low permeability formations (B3 and B5) could be explained by the specific storage (compressibilty). Recall from Chapter 3.1.4 that rapid pore pressure changes within aquitards could be explained with the Terzaghi equation (Equation 3.2) by the coefficient of consolidation which is equivalent to hydraulic diffusivity. This parameter then defines the relationship between hydraulic conductivity and specific storage.

A hydraulic diffusivity of about 10^{-3}, and a Ss of $\sim 5\times10^{-4}$, was required to achieve the observed drawdown within the lower aquitard. The calibrated Ss is comparable to the maximum Ss of 3.5×10^{-4} which was measured on minimally disturbed core of this aquitard.

6.4.2 Sensitivity to aquitard window width

The multi-layered flow model was a useful tool with which to test the effects of aquitard windows on vertical flux. In reality, aquitard windows are more likely to be disc or irregular in shape, however, the axisymmetric nature of the model meant that aquitard windows were ring shaped, centered on the irrigation bore. Despite this drawback, the proportional of aquitard window to the total area of the aquitard layer was able to varied to proportions likely to be realistic. Table 6.6 for example, shows that for windows at a distance of 1400 m from the irrigation bore, widths between 24–120 m corresponded to total 'holeyness' of between 2.2–11.2%.

The proportion of vertical leakage to lateral flow in the deep aquifer was sensitive to the inclusion of an aquitard window and in particular, window width and proximity to the pumping bore. In the absence of an aquitard window, approximately 90% of inflow was lateral and 10% vertical leakage through the silt unit. For a window ~50 m wide and 1 400 m from the pumping bore, the proportion of inflow from vertical leakage through the aquitard was approximately 8%, vertical leakage through the aquitard window was

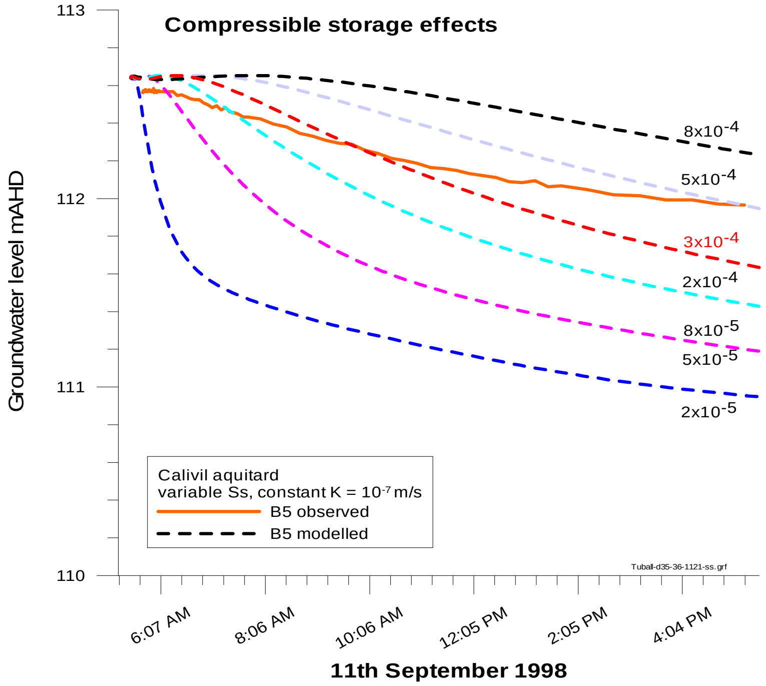


Figure 6.14: Sensitivity of aquitard drawdown to compressible storage with no aquitard windows (K $= 10^{-7}$ m/s).

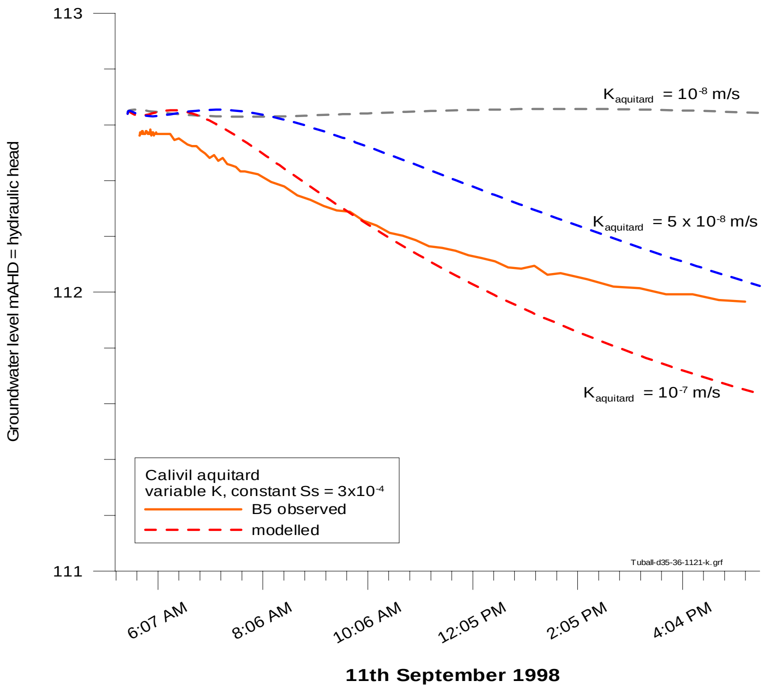


Figure 6.15: Sensitivity of aquitard drawdown to permeability including aquitard windows (Ss $= 3\text{x}10^{-4}$).

Table 6.6: Examples of the relationship between aquitard window width and the proportion of vertical and lateral flow.

Model ID	Calivil aquitard K (m/s)	Window location radius (m)	Window width (m)	Area window versus aquitard %	Q thru window %	Q thru aquitard %	Q lateral %
15.03a	1.00E-09	no win	-	-	-	1.0	99.0
15.03b	1.00E-07	no win	-	-	-	10.8	89.2
15.03c	1.00E-06	no win	-	-	-	20.4	79.6
15.03	1.00E-09	1400	24	2.2	1.5	28.0	70.6
15.04a	1.00E-07	1400	47	4.3	17.9	9.2	73.0
15.04b	1.00E-08	1400	47	4.3	23.4	4.5	72.1
15.04	1.00E-09	1400	47	4.3	21.7	8.3	70.0
15.05	1.00E-09	1400	70	6.4	22.6	6.3	71.1
15.06	1.00E-09	1400	120	11.2	23.8	5.2	71.0

22% and lateral flow across the boundary was reduced to 70%, though these results were dependent on the permeability of the aquitard.

It was found, in this groundwater system, that aquitard windows smaller than ~20 m may significantly change the flow balance, but that above a threshold width, little increase in vertical flux occurred. This principle is depicted by Figure 6.16, where the threshold window width for an aquitard of $K_v = 10^{-9}$ m/s was <20 m, but was ~70 m for an aquitard of $K_v = 10^{-7}$ m/s . Where an aquitard window is smaller than the threshold width for the given aquitard-aquifer system, vertical flux is dominated by leakage through the aquitard itself. The insensitivity of flow to window length above a threashold ratio agrees with theoretical work by Strack (1981a) and Strack (1981b).

The significance of leakage through an aquitard window was reflected by high groundwater velocity (Figure 6.17). A cross-section through the lower aquitard (layer 6) is typical of the scenarios which were modelled, with groundwater velocity highly sensitive to the existence of a window of higher permeability, in contrast to the insensitivity of hydraulic head. Over time, however, sufficient flux may occur to measurably impact groundwater levels in inter-connected aquifers.

Nevertheless, this result implies that calibration of hydraulic head does not guarantee that a flow model realistically describes vertical fluxes that may be of significance for solute transport. While Fogg (1986) demonstrated that calibration of a regional flow model to observed hydraulic head was insensitive to spatial variability of permeability, these results suggest that a localised, relatively constrained flow model may be similarly limited.

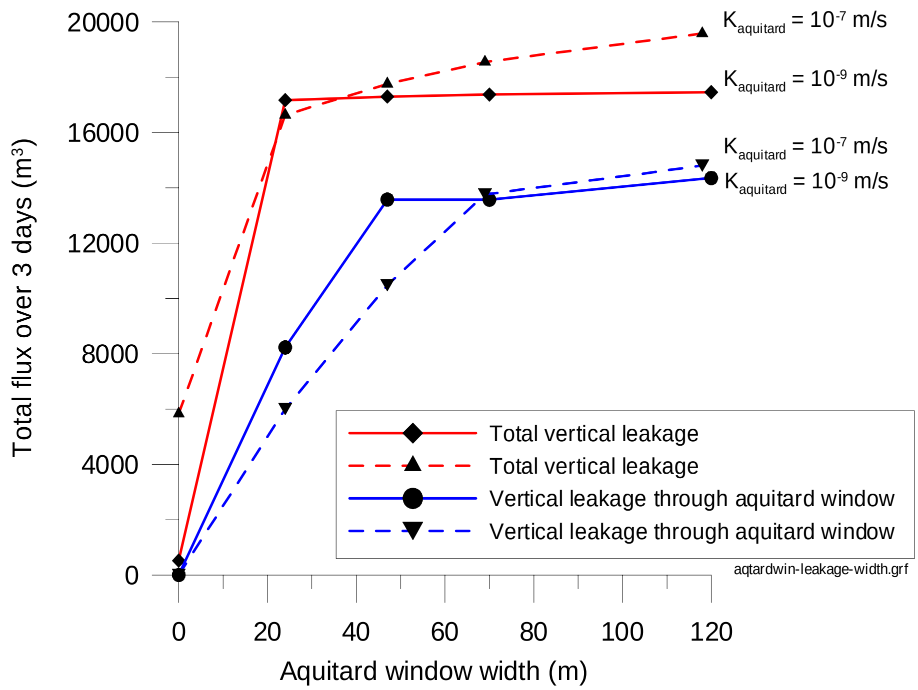


Figure 6.16: Vertical flux related to aquitard window width, showing narrow windows greatly increase vertical flux and a threashold width beyond which little increase in vertical flux occurs.

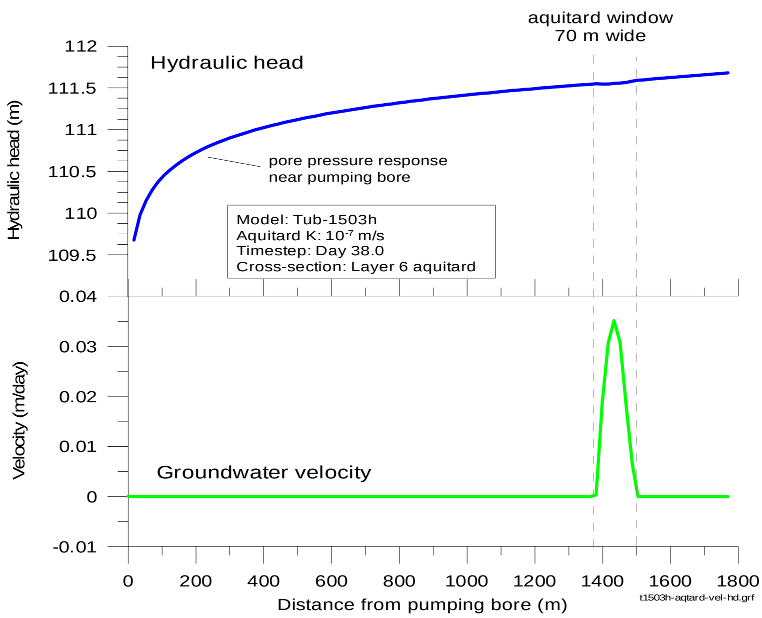


Figure 6.17: Cross-sections through layer 6 aquitard, showing high sensitivity of groundwater velocity, and insensitivity of hydraulic head to a 70m wide window.

Table 6.7: Vertical leakage via boreholes.

Model ID	Radius (m)	Annulus K (m/day)	Annulus leakage %	Aquitard leakage %	Lateral leakage %
30.me	0	-	-	19.7	80.3
30.03	0.2	40	0.035	19.7	80.3
30.04	0.2	100	0.084	20.1	79.8
30.02	0.4	40	0.24	19.1	80.6
30.02a	0.4	100	0.58	-	-
30.02b	0.4	200	1.16	-	-

6.4.3 Sensitivity to bore casing leakage

Vertical leakage via bore casings was generally of secondary importance to large scale rapid flow pathways. Given the lesser importance of bore casing leakage, and time constraints on execution of these lengthy model scenarios, a limited number of model scenarios was attempted.

Using the high density, regular rectangular mesh, which has been described, vertical leakage through a 0.2 and 0.4 m radius leaky borehole-gravel pack was estimated. Leakage through a 0.2 m width ring shapped annulus could then be estimated by difference. The hydraulic conductivity was set at 40 and 100 m/day, equivalent to the likely range for a gravel packed borehole. It should be noted, however, that the permeability of an open borehole, as may occur for a corroded casing, is infinite and would greatly increase leakage estimates.

The proportion of leakage to the deep aquifer was very small (<1%), and had no significant impact on other flow pathways (Table 6.7). However leaky boreholes are likely to provide an important 'short-circuit' between the shallow and middle aquifer. The spatially extensive indurated clayey sand forms an effective hydraulic barrier, resulting in a maximum gradient of 0.67 which constitutes a large driving force for leakage (Lacombe et al. 1995). Despite the small proportion of vertical leakage, a large concentration difference between the shallow and middle aquifers may result in significant groundwater quality impact, particularly given an unknown number of leaky boreholes in the area.

6.5 Model limitations

Upper Shepparton aquifer

A recharge boundary was included in some models, but was of little added benefit due to the lack of hydraulic constraints on flow in the Upper Shepparton aquifer. Downwards flow from layer 1, was controlled by known aquitard permeability, however, the gradient and permeability of the aquifer were unknown. Some recharge is likely to have occurred during the modelling period due to a rainfall event of 27 mm on the 12th September and 12.4 and 13.8 mm on the 23-24th September. Although significant time delays are probable, given the 20 m thickness of the semi-saturated zone, increased hydraulic pressure is likely to be transmitted to shallow groundwater, particularly in more permeable areas. Annual average recharge for these cells in the regional model of Punthakey et al. (1994) ranged between negative values (excessive evaporation) and 365 mm/year.

Seasonal drawdown within the Upper Shepparton aquifer was not observed during the 15 day period that was modelled. This meant that vertical leakage estimates relied solely on measured K for the indurated clayey sand. If the entire irrigation season had been modelled, simulated drawdown could have been compared with observed drawdown of 0.2 m. However, interference effects on nearby bores would complicate interpretation for the extended period.

Axisymmeteric flow

Since this is an alluvial fan deposit, it is highly unlikely that axisymmetric flow will occur. A full 3D model, or non-axisymmetric modelling approach such as that developed by Barrash & Dougherty (1997) would be more suitable. However, since there are only 2 nested piezometer sites with stratigraphic data available, our understanding of the system would not be improved and this approach is therefore unwarranted.

6.6 Summary

- Groundwater level changes in specific aquifer and aquitards at the Tubbo site during the 1998-1999 irrigation season were described by analytical pump test analysis and numerical flow modelling. The largest vertical gradients developed during pumping over the indurated clayey sand between the shallow and middle aquifers. The gradient over a specific aquitard thickness was larger than aquifer averaged gradient.

- Direct flow via aquitard windows was indicated by drawdown and recovery in aquifers prior to intervening aquitards, and other hydraulic evidence. Aquitard discontinuities seriously limited drawdown analysis by analytical methods.

- Calibration of an axisymmetric flow model, centered on the irrigation bore could be achieved with either spatially continuous or discontinuous aquitards, despite constraints by measured hydraulic parameters and transient flow boundaries. During pumping, lateral fluxes dominated, but large vertical leakage occurred from the middle aquifer. Volumetric leakage from the shallow to middle aquifer was relatively minor.

- Aquitard windows narrower than 20 m could greatly increase vertical flux, but no additional increase occurred for windows wider than a threshold value. Hydraulic heads were insensitive to large groundwater velocity within a localised areas implying that a model calibrated to observed groundwater levels does not guarantee the absence of aquitard windows.

Chapter 7

HYDROCHEMICAL VARIABILITY

In this chapter, vertical leakage is detected using hydrochemical and isotopic techniques. Intensive monitoring of hydrochemical changes due to downwards flux is particularly useful at depths beyond the reach of surface geophysical techniques, and where there are few boreholes to delineate subsurface stratigraphy. Evidence for mixing of groundwater with different hydrochemical signatures over the medium term, and during the 1998-1999 irrigation season, is presented and quantified for the shallow aquifer with hydrochemical mass balance modelling.

7.1 Results

Groundwater was sampled from 40 piezometers at 7 selected sites located in the Lower Murrumbidgee area (Figure 7.1). Piezometers at the Tubbo site, were sampled intensively, 6 times between Aug-98 and May-99. Two additional sites for surface water sampling included the Murrumbidgee River at Euroley Bridge and the groundwater fed irrigation channel at Tubbo. Datasets produced for groundwater, surfacewater and rainfall hydrochemistry, and historic data obtained from DLWC files are outlined in Table 7.1.

7.1.1 Unstable parameters

Electrical conductivity (EC) of groundwater varied considerably with depth, relative to the aquifer-aquitard matrix and also changed considerably during the irrigation season. Salinity at selected sites in Feb-99, varied from fresh (148 μS/cm) to saline (11 490 μS/cm), with an average of 1986 μS/cm and median value of 442 μS/cm. At the Tubbo site the most dramatic change occurred below the indurated clayey sand, where a maximum EC of >1 000 μS/cm in Sep-98 decreased to <400 μS/cm by Feb-99 (Figure 7.2). A large EC increase was observed in aquifer 1, and a minor decrease observed in aquifer 3. The lowest

Table 7.1: Hydrochemical data used in the study.

Site details	Sample date	N [a]	Sample ID	Comments
DLWC data	1972–1995	44	0.01-0.45	Limited parameters
Tubbo site	Aug-98	6	0.91-0.96	Unstable parameters only
Tubbo site	Sep-98	9	1.01-1.10	
Tubbo site	Oct-98	10	2.01-2.11	
Tubbo site	Dec-98	8	3.01-3.10	
Tubbo site	Feb-99	10	4.17-4.25	
Selected sites	Feb-99	23	4.01-4.17, 4.26-4.34	
Tubbo site	May-99	9	5.01-5.10	
Selected sites	Feb-01	10	15.05-15.11, 15.22 -15.25	
	Total	85		
Leeton rainfall	Sep-99 to Aug-00	10	12.01-12.19	Some samples not analysed

[a]N is no. of samples

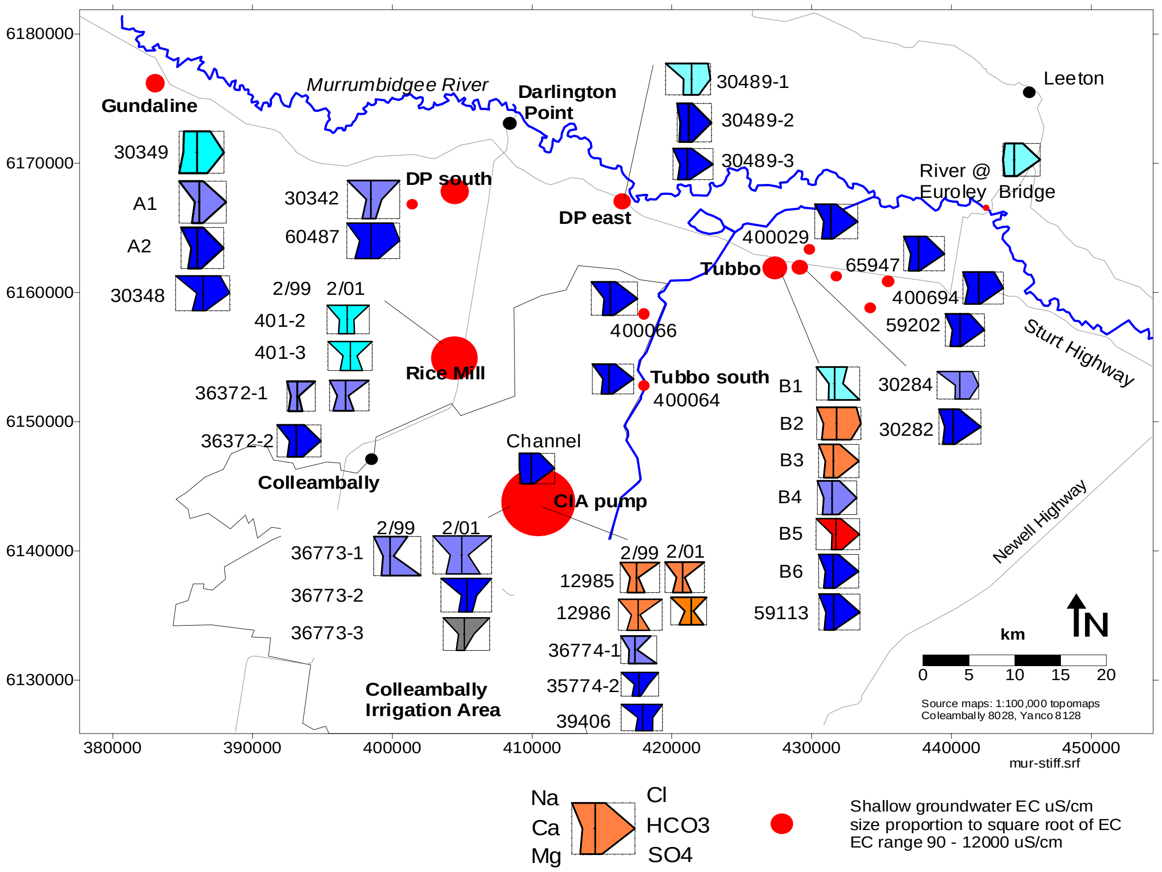


Figure 7.1: Spatial variability of salinity and watertype, February 1999 & 2001.

EC measurements were in the middle silt unit and aquifer 4.

Regional groundwater is near neutral pH (6.2-7.7). At the Tubbo site, during the irrigation season pH varied between 5.5-8. The depth profile of pH was most variable in Sep-98 and Oct-98. By Feb-99, pH profiles had become more homogenous with depth, with pH about 6.5, with slightly high pH at aquifer 1 and the lower silt unit.

Regional groundwater was generally oxidising, characterised by dissolved oxygen (DO) of 0.1-9.2 mg/L (mean 3.2 mg/L). At Tubbo, DO and Eh generally increased during the irrigation season at all depths. DO was highest in the middle silt unit, reaching a maximum of 10.5 mg/L in May-99 and was consistently low in the lower silt unit. The largest increase in both Eh and DO during the irrigation season occurred in the middle silt unit.

Groundwater temperature ranged from 19.0-26.3 ^{0}C, with a mean of 23.0 ^{0}C (except for irrigation bores which are possibly warmed by pumping activity). Groundwater temperatures increased with depth at the rate of 3.8 ^{0}C per 100 m. At Tubbo, temperature generally increased with depth and decreased during pumping. Water sampled from the irrigation turbine pump showed the greatest variation over time, with some high temperature anomalies. Temperature in aquifer 1 and 2 increased by the end of pumping in May-99.

EC variability during the past 20 years

Salinity trends in the Shepparton aquifer at selected sites over the past 20–25 years, were variable (Table 7.2, Figure 7.3). No corresponding trends were evident in the Calivil aquifer. Decreasing and stable EC values at the Gundaline and Tubbo sites respectively, contrasted with increasing EC values in shallow aquifers over the past 5 years at sites within and to the north of the CIA.

The most rapid rates of increase occurred between 1986 and 1999 at North Coleambally (36372-1, 47 m depth) where EC increased from <500 to 4 120 μS/cm. This was equivalent to an average increase of total dissolved salts (TDS) of 133 mg/L/year. At Darlington Point south (30342, 50.7 m depth), EC values rose two fold between 1980 and 1999, from <600 to 1 260 μS/cm.

7.1.2 Major ions

Spatial variability - selected sites

Change in watertype highlights the change in relative proportion of major ions. The Szczukariew-Priklonski classification (Alekin 1970) of groundwaters is adopted here due

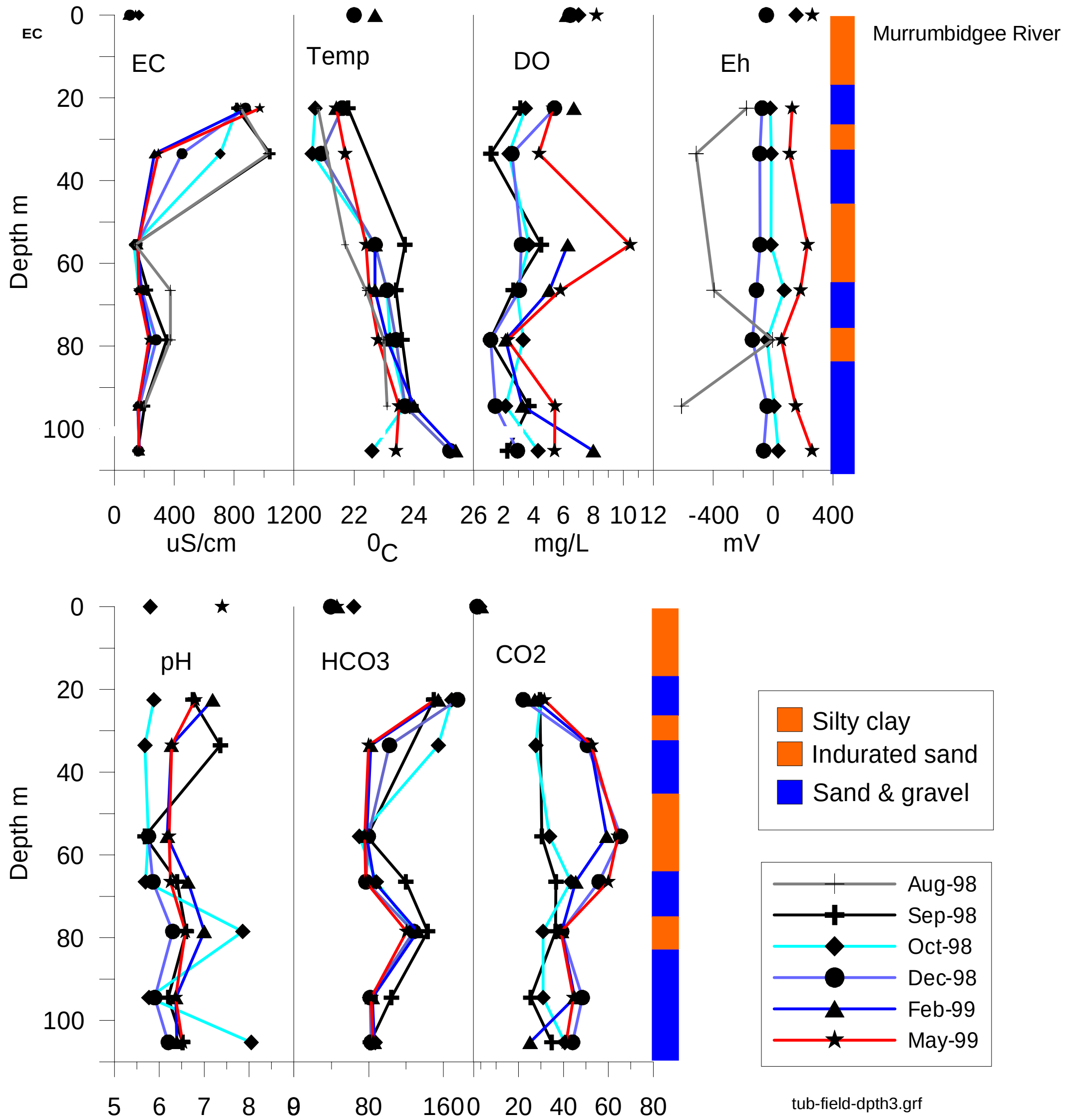


Figure 7.2: Variability of unstable parameters with depth at Tubbo site, 1998/99.

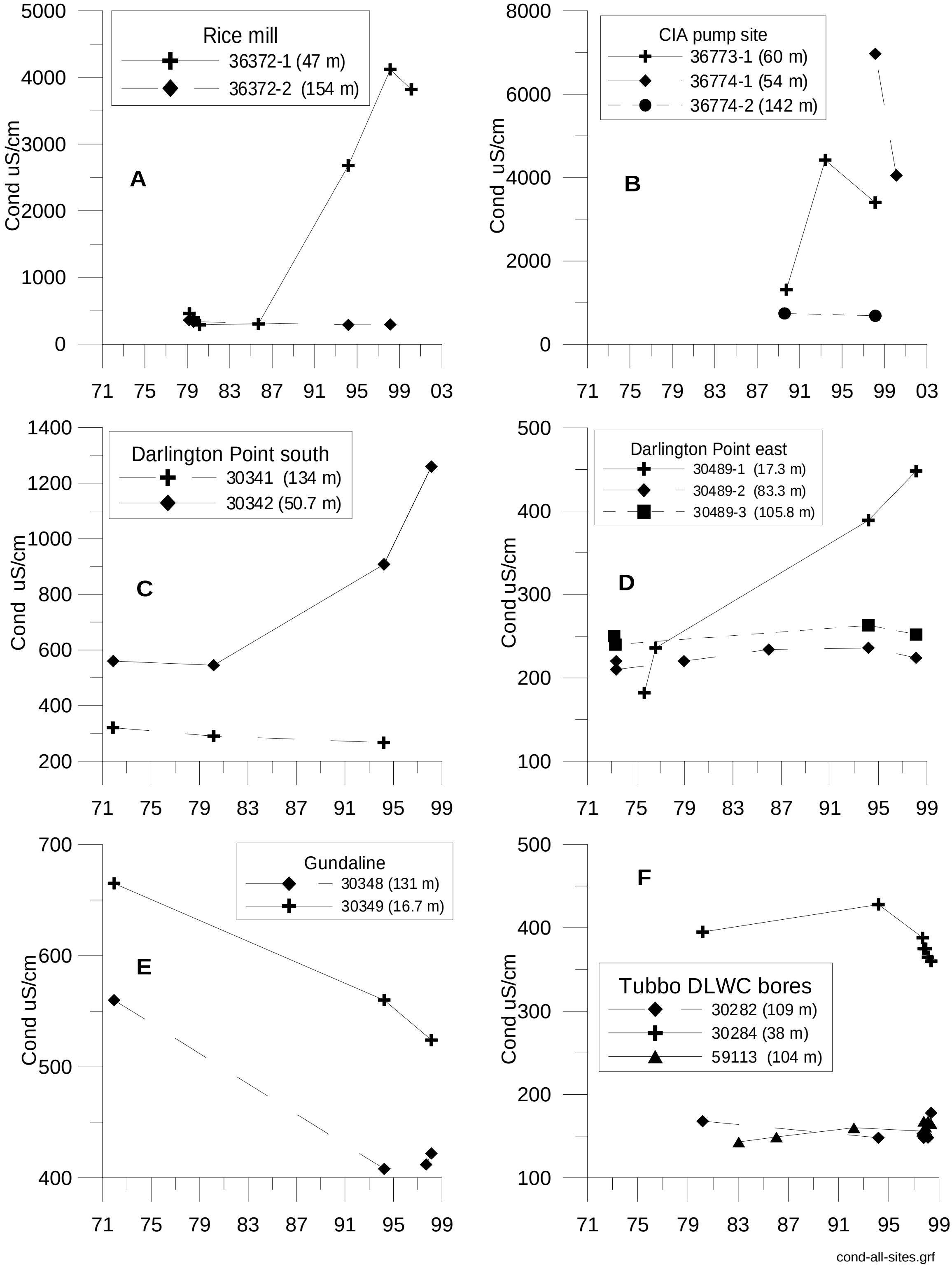


Figure 7.3: Salinity variability in shallow and deep aquifers at selected sites, 1970–1999 **A** Rice Mill site **B** CIA pump site, **C** South Darlington Point site, **D** Darlington Point east site, **E** Gundaline site, **F** Tubbo site.

Table 7.2: Salinity trends at nested piezometer sites, pre-1980–1999.

Site	Trend	Pre 1980 EC (μS/cm)	1999 EC (μS/cm)
Darlington Point south (30342)	Increasing	<600	1 260
Darlington Point east (30489-1)	Increasing	<250	448
Rice Mill (36372-1)	Increasing	<500	4 120
CIA pump (36773-1)	Variable	1 310	4 050
Gundaline (30348)	Decreasing	<700	524
Tubbo (30284)	No trend	~400	365

to its comprehensive classificaton of water type best suited to salinity studies. This classification scheme is based on the occurrence of 6 major species (combining Na+K), which comprise more than 20% of total anions and total cations respectively, calculated in meq/L.

Shallow groundwater within clayey-silt is typically Na-Mg-Cl-SO_4 (eg. B1, 12985). Deep groundwater, which is abstracted for irrigation is typically Na-Mg-Ca-HCO_3-Cl, except at the CIA pump site (39406) and Darlington Point south site (60487) which are of the Na-Cl-HCO_3-SO_4 type.

Water from the Murrumbidgee River was of the Ca-Mg-Na-HCO_3-Cl-SO_4 type. This contrasts with Irrigation channel water (at CIA pump) and shallow groundwater at the Gundaline site (30349) which was of the Mg-Na-Ca-HCO_3-Cl type.

Variability during the past 30 years

Major ion variability was examined between 1970 and 2001, based on available hydrochemical data obtained from DLWC archives.

Of 6 sites, significant medium term hydrochemical changes were identified at 2 sites, minor changes at 3 and no change at the CIA pump site over the short period of records since 1990. These changes are evident in Piper diagrams (Figure 7.4). Darlington Point south was the only location where hydrochemical changes occurred in both the middle (51 m) and lower (134 m) aquifers. Between 1972 and 1999, water types evolved to more Cl rich waters, while cation proportions remained constant except for an increase in Na by 1999 in the middle aquifer. Minor changes occurred in the shallow (17 m) aquifer at Darlington Point east site, with an increase in both Na and Cl. By contrast, at the Gundaline site, the proportion of Na and Cl decreased in the shallow aquifer between 1972 and 1999. At the Tubbo site, a small relative increase in Na was associated with a

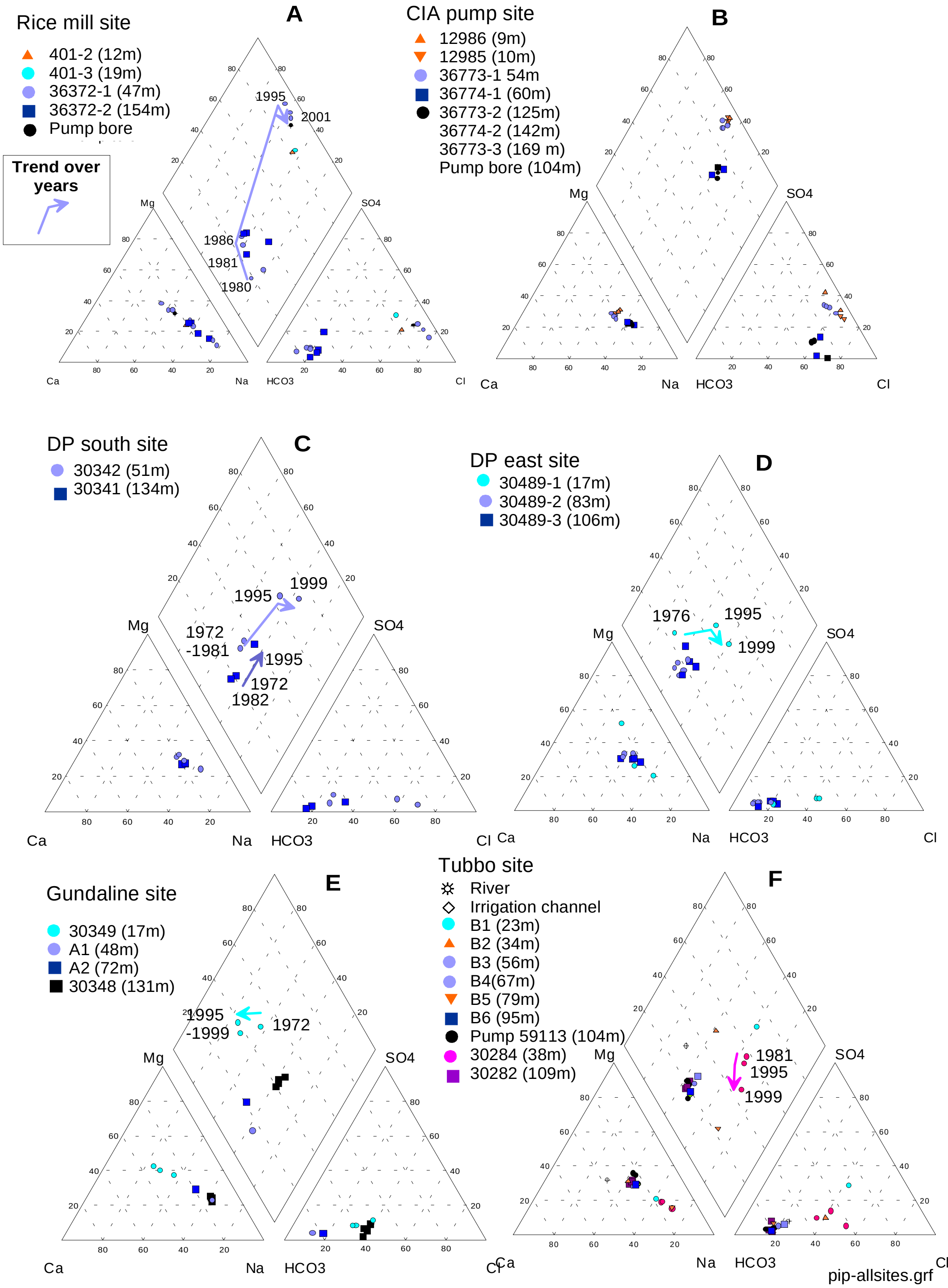


Figure 7.4: Piper diagrams showing hydrochemical variation at selected sites 1970–2001, **A**-**D** Salinisation of shallow aquifer **E**-**F** Freshening of shallow aquifer.

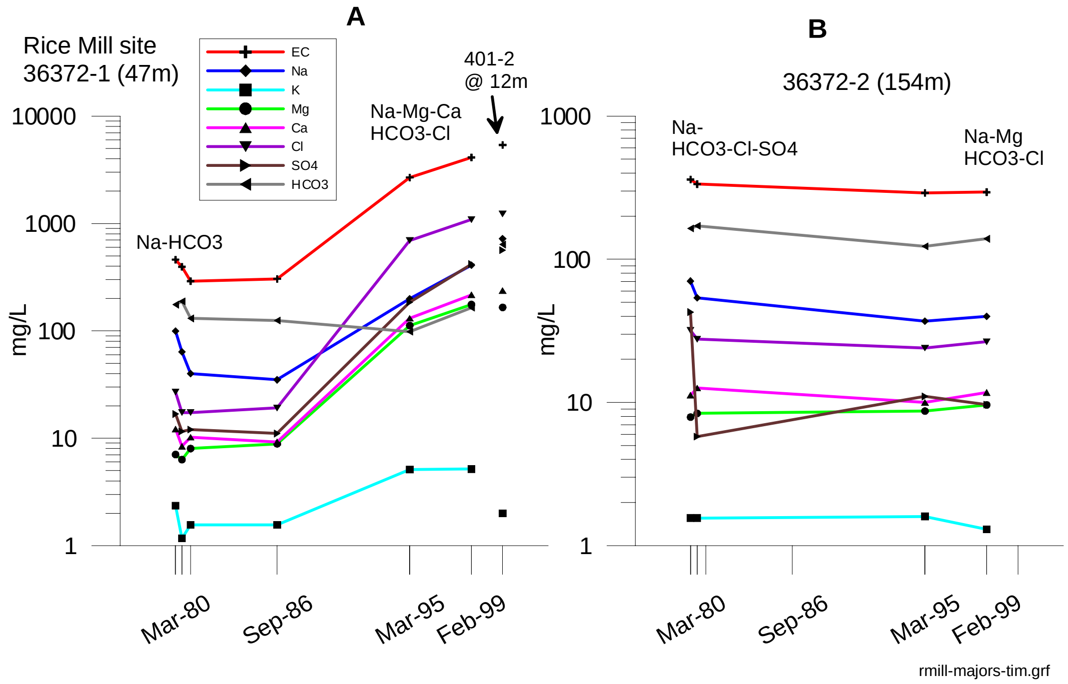


Figure 7.5: Major ion variability at Rice Mill site, 1980–1999 **A** 36372-1 **B** 36372-2.

decrease in Cl.

At three of the sites, salinisation occurred in two stages. Firstly, steady Na or increase in Ca, and then after 1995, an increase in Na. The changing trend direction on the piper diagrams is clearly evident in Figure 7.4.

The most significant hydrochemical changes occurred at the Rice Mill site where Na-HCO_3 water at 47 m evolved to a Na-Mg-Ca-Cl-SO_4 type water between 1980 and Feb-99. Major ion trends at the Rice Mill site between 1980 and 1999 were examined in detail, as hydrochemistry was sampled 6 times over this period (Figure 7.5).

Variability during the 1998-1999 irrigation season at Tubbo site

The concentration of all major ions is highest in aquifer 1 and within the clayey cemented sand in Sep-98 and Oct-98 (Figure 7.6). All major ion concentrations decrease with depth, except for a large Na increase and small increase in other ions within the lower silt unit. During the irrigation season, the concentration of all major ions decreased within the lower silt unit and increased in aquifer 1. However, in aquifer 1, Ca and Mg concentrations decreased, and Na increased between Sep-98 and Oct-98. During the remainder of the irrigation season, Ca and Mg concentrations increased while Na decreased. The largest

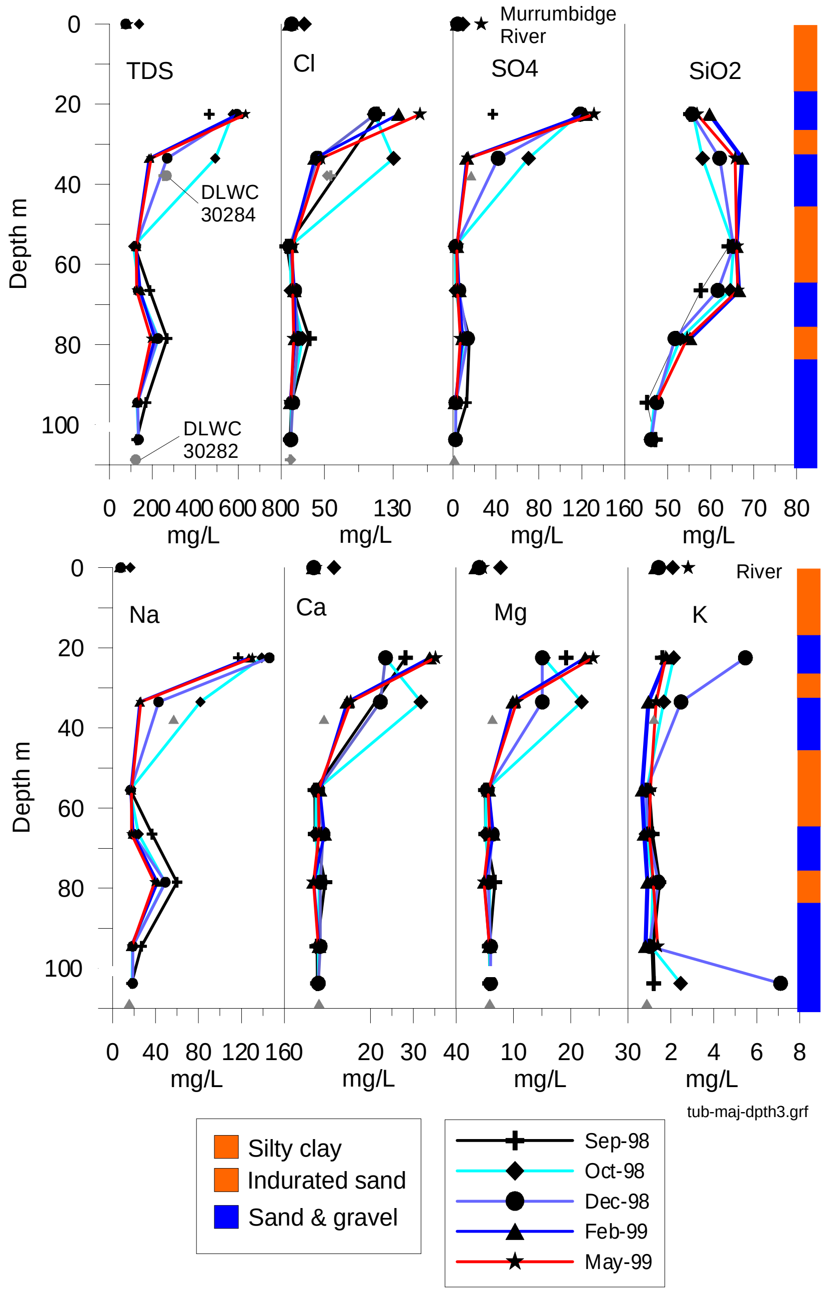


Figure 7.6: Depth profiles of major species at Tubbo site, 1998–1999.

changes in concentration occurred in aquifer 2 where all major ion levels decreased.

Small hydrochemical changes occurred in aquifer 3 and 4 which are more distinct on temporal rather than depth plots (Figure 7.7). Between Oct-98 and Dec-98, Cl increased in aquifer 3 (B4) from 11.1 to 15.4 mg/L, a gain of 38.7%. During this time Cl increased in aquifer 4 (B6) from 11.6 to 13.1 mg/L, a gain of 12.9%. Other major ions also increased including Ca, Mg and SO_4, although HCO_3 and Na levels were constant or slightly decreased. In aquifer 3, Cl concentrations had decreased but not returned to the concentration prior to pumping during the last sampling in May-99. In aquifer 4, Cl concentrations had decreased to pre-pumping levels by Feb-99.

Spatial changes can be evaluated by comparing chemistry at the nested piezometers (B4 and B6) within 40 m of the pumping bore and at observation bores 1.8 km distant (30284 and 30282). Major ion concentrations in 30284 were relatively constant, except for a small constant decrease in Cl from 57.6 to 48.9 mg/L (-15%) between Sep-98 and Feb-99. Comparing major ion concentrations of B2 and 30284 in May-99, anion concentrations were similar, Na was higher in 30284 while Mg and Ca levels were lower. At 30282, no significant changes in Cl or Na were observed, though decreased SO_4 and K occurred, together with a small increased in Ca and Mg during Feb-99.

Variance, σ^2, is a measure of scatter of values around a mean (Swan & Sandilands 1995). The σ^2 of Cl concentration in B6 was much greater than the variance in the same aquifer, but 1.8 km away from the pumping bore (at 30282). The first Cl value for B6, was made equal to the 2nd measurement in order to eliminate the probable effects of drilling fluid. Based on 6 measurements over a 7 month period (Sep-98 to May-99), σ^2 was calculated to be 1.13 and 0.25 at B6 and 30282 respectively. These values indicate that water quality variation was greatest near the irrigation bore (B6).

The association between groundwater quality and pumping activity is also demonstrated by the timing of the chloride peak in B6, which occurred in the first sample after pumping began.

Changing watertypes during 1998–1999 irrigation season at Tubbo site

Watertypes changed over the short term at Tubbo site. Table 7.3 shows the initial and final watertypes by the Szczukariew-Priklonski classification. The largest changes occurred in aquifer 1 with the increasing dominance of SO_4. Major changes also were observed in aquifer 3 which evolved from a distinctive Na-HCO_3 water to a watertype similar to other aquifers. In the Lower silt unit, Na and HCO_3 became increasingly dominant.

Temporal variability of major ions in the shallow aquifers (1&2) is distinct from

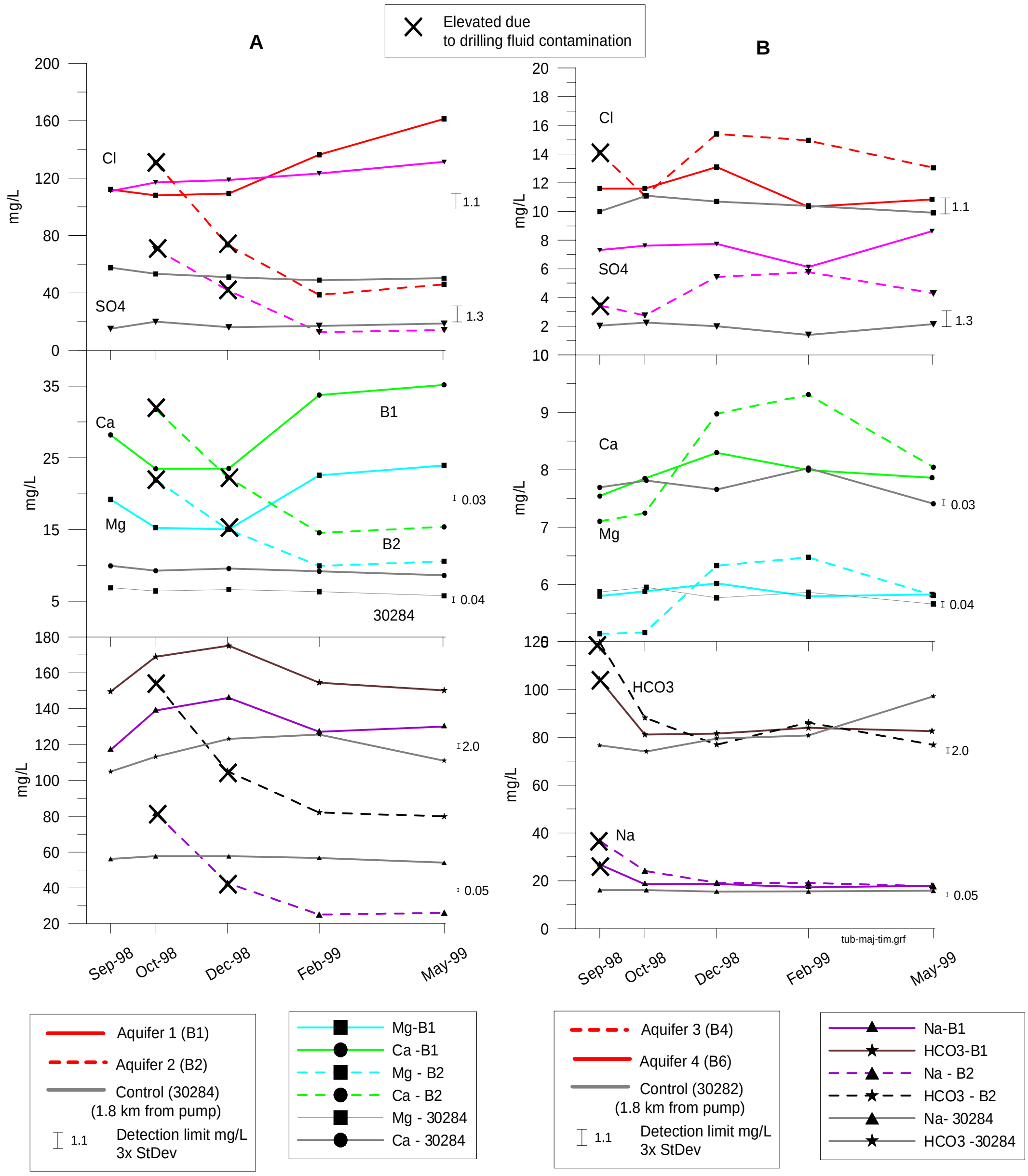


Figure 7.7: Major ion variability at Tubbo site, Sep-98 to May-99 **A** Aquifer 1 & 2 **B** Aquifer 3 & 4.

Table 7.3: Initial and final watertypes for surface and groundwaters at Tubbo, Sep-98 and May-99.

Aquifer/aquitard	Initial watertype Sep-98	SP [a] class	Final watertype May-99	SP class
River	Ca-Mg-Na-HCO_3-Cl-SO_4 (Dec-98)	219	Na-Mg-Ca-HCO_3-Cl	142
Channel	Na-Mg-Ca-HCO_3 (Oct-98)	58		
Aquifer 1	Na-Cl-HCO_3-SO_4	65	Na-SO_4-Cl	30
Clayey sand	Na-SO_4-Cl-HCO_3	66	Na-Mg-Ca-HCO_3-Cl-SO_4	214
Aquifer 2	Na-HCO_3-Cl-SO_4	68	no change	
L.Shep aquitard	Na-Mg-Ca-HCO_3	58	Na-Mg-Ca-HCO_3-Cl	142
Aquifer 3	Na-HCO_3	7	Na-Mg-Ca-HCO_3-Cl	142
Calivil aquitard	Na-HCO_3-SO_4-Cl	69	Na-HCO_3	7
Aquifer 4 (B6)	Na-Mg-HCO_3	22	Na-Mg-Ca-HCO_3	58
Aquifer 4 (30282)	Na-Mg-Ca-HCO_3	58	Na-Mg-HCO_3	22
Aquifer 4 (59113)	Na-Mg-Ca-HCO_3	58	no change	

[a]SP = Szczukariew-Priklonski classification (Alekin, 1970).

the trends in the deep aquifers (3&4). Significant increases, above detection limits were observed in Ca, Mg, Cl and SO_4 concentrations. A sustained steady increase (+67%) in Cl (120–160 mg/L) occurred in aquifer 1, between Dec-98 and May-99. Changes in the deep aquifers occurred more quickly after the start of pumping, but to a lesser degree. In aquifer 3, Cl increase (+47%) occurred between Oct-98 and Dec-98, with a decline by May-99. In aquifer 4, Cl increase (+30%) occurred in the same period, but had returned to pre-pumping levels by Dec-99. No significant hydrochemical changes were observed in the DLWC observation bores, 1.8 km distance from the pumping bore.

No corresponding Na and HCO_3 trends were observed, due to the similar concentrations in each aquifer. Contamination by drilling fluid is clearly evident between Sep-98 and Oct-98 in all aquifers near the observation bore.

7.1.3 Minor and trace elements

Nitrate

Nutrient (NO_3, NO_2, NH_4, PO_4) results for ground and surface waters have implications for water quality management, but will not be discussed here, except to highlight nitrate contamination that may be related to vertical leakage. The full nutrient dataset is included on the accompanying CD-ROM.

Nitrate was detected in all bores and piezometers sampled during Feb-99, at concentrations ranging from 0.2–9.0 mg/L NO_3-N, with a mean of 0.4 mg/L NO_3-N. A maximum concentration of 9 mg/L NO_3-N was detected at observation bore 30489-1. Concentrations >2 mg/L NO_3-N were detected in observation bores 36372-1, 36773-1, 36774-1, and shallow piezometer 12985.

At the Tubbo site between Sep-98 and Feb-99, nitrate levels ranged between 0–1.4 mg/L NO_3-N. Nitrate concentrations were highest in Dec-98, with maximum concentration in aquifer 1 and generally decreased with depth.

Iron

Detailed Fe^{2+} and FeT results are available on the accompanying CD-ROM. Of relevance here are high total iron levels reported for some bores which suggest that corroded bore casings may contribute to vertical leakage.

Maximum iron concentrations of 2 mg/L were found in the indurated clayey sand and lower silt unit. Higher iron concentrations of up to 20 mg/L FeT were detected in the steel cased observation bores (30282 & 30284). High (66.0 mg/L) iron concentration was detected in bore 30342, which was attributed to corrosion of the steel casing.

Table 7.4: Strontium and manganese concentrations detected, Feb-99 and Feb-01.

	Sr (mg/L)	Mn (mg/L)
No. of detections	44	21
No. of samples	44	44
Detection Limit	0.03	0.01
Max	9.90	2.13
Mean	1.16	0.33
Median	0.20	0.16

Trace element chemistry

Variable concentration with depth for three of the most common trace elements, Sr, Li and Mn is depicted in Figure 7.8. The indurated sand layer and other aquitards appear to be sources of Sr and Mn.

Strontium (Sr), was the most common trace element detected, ranging from 0.08–9.9 mg/L, followed by manganese (Mn) which ranged from 0.01 to 2.13 mg/L (Table 7.4). Strontium concentration decreased with depth at the Tubbo site from 0.39 mg/L in B1 to 0.16 mg/L in B2, and <0.1 mg/L at greater depth. During the irrigation season, strontium concentrations increased from 0.3 to 0.39 mg/L in the shallow aquifer (B1)

Trace concentrations of copper (Cu), were detected in riverwater and several groundwater bores, while lithium (Li) and boron (B) were detected in shallow groundwater within, or directly below clayey silt deposits (Table 7.5).

Other trace elements were not detected in the mg/L concentration range: cadmium, chromium, nickel, lead, and zinc.

Table 7.5: Copper, lithium and boron trace concentrations detected, Feb-99.

	Cu (mg/L)	Li (mg/L)	B (mg/L)
No. of detections	4	4	2
No. of samples	34	34	34
Detection Limit	0.006	0.006	0.021
River	0.0104	BDL	BDL
30349	0.056	BDL	BDL
60487	0.0085	BDL	BDL
59113	0.0099	BDL	BDL
12985	BDL	0.046	1.055
12986	BDL	0.064	0.499
36774-1	BDL	0.02	BDL
36773-1	BDL	0.01	BDL

BDL is below detection limit

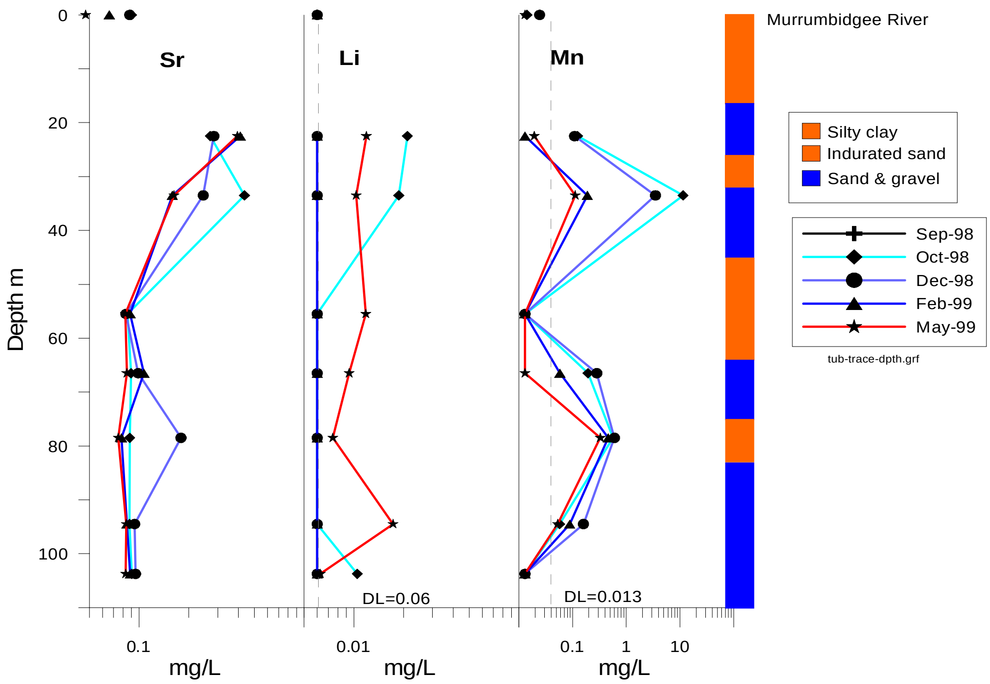


Figure 7.8: Trace element variability with depth at Tubbo site, Sep-98 to May-99.

7.1.4 Stable isotopes

Environmental isotopes, elements with varying atomic mass, now routinely contribute to groundwater investigations, providing a unique insight into recharge processes, groundwater quality, rock-water interaction, and the origin of salinity. (Clark & Fritz 1997). Stable isotopes such as carbon-13 (δ^{13}C), deuterium (δ^{2}H) and oxygen-18 (δ^{18}O), which are used in this study, provide a natural signature tracer for groundwater origin.

Conventions for reporting of isotope ratios, fractionation and enrichment have been introduced in Chapter 4.3, and the full isotope data set is included on the accompanying CD-ROM.

Establishing a local meteoric water line

A local meteoric water line (LMWL) was established for Leeton (Figure 7.9). Determining a LMWL was important to understanding the values of ^{18}O and ^{2}H observed in ground and surface waters, relative to modern rainfall. For instance, some of these samples plot to the left of the GMWL, but to the right of the Leeton LMWL. This distinction is important in interpreting the effects of mixing versus other geochemical processes such as silicate hydration.

The Leeton LMWL was defined as follows:

$$\delta^2 H = 7.3\delta^{18}O + 11.3 \tag{7.1}$$

The LMWL was based on 10 samples collected from rainfall events between Sep-99 and Aug-00 (S. Lawson, pers. com.) at the township of Leeton, about 20 km northeast of the Tubbo site. Samples were collected by the method outlined in Chapter 4.5.3, and samples from events <6 mm magnitude were discarded. This was to avoid the enriching effects of secondary evaporation during passage of raindrops through a warm, dry atmosphere typical of semi-arid areas - a process known as the amount effect (Dansgaard 1964).

The LMWL for Leeton is similar to the global meteoric water line (GMWL), that was first defined by Craig (1961) and later refined by Rozanski, Araguas-Araguas, & Gonfiantini (1993) (Equation 7.2). The lower slope may be attributed to evaporation during rainfall. Identical ^{2}H intercept is a coincidence which is controlled by primary evaporation in the source region of water vapour (Clark & Fritz 1997). The mean $\delta^2 H$ was -22.67 ‰ and the mean δ^{18}O -4.67 ‰ .

$$\delta^2 H = 8.17\ \delta^{18}O + 11.3 \tag{7.2}$$

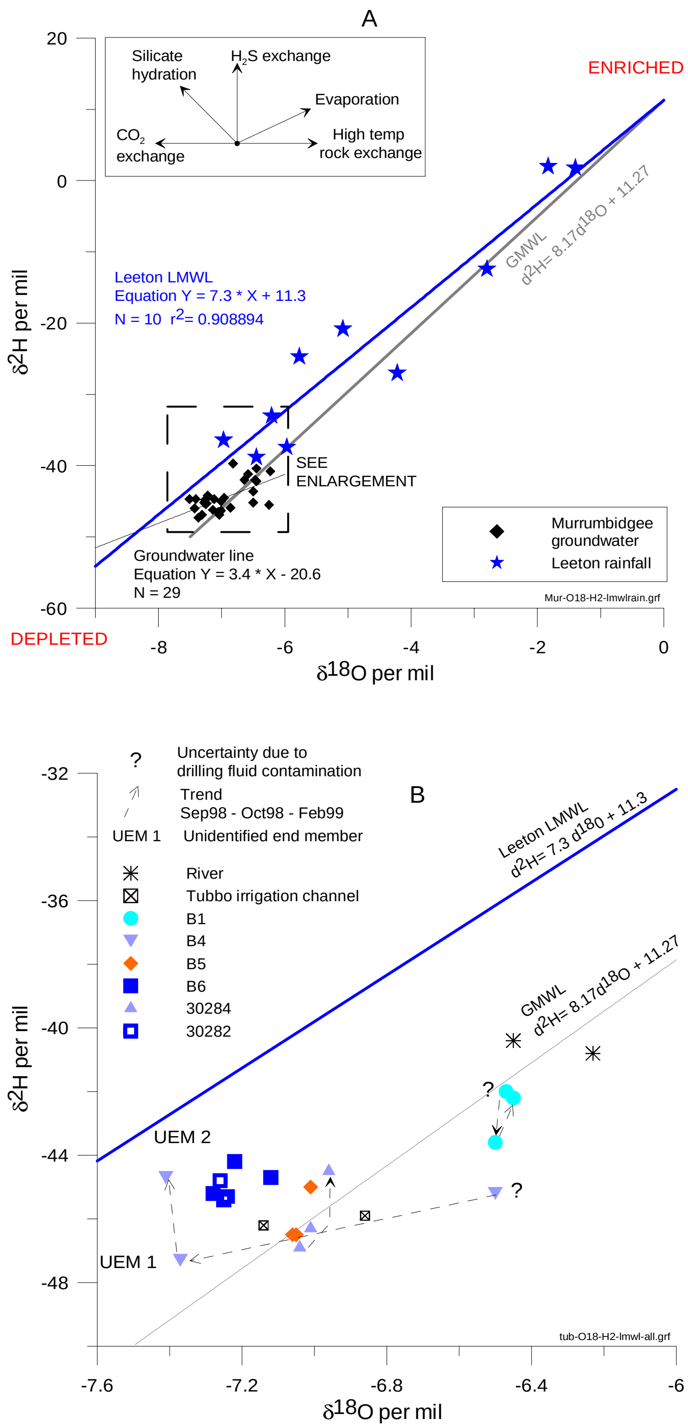


Figure 7.9: Oxygen-18 and deuterium data from Lower Murrumbidgee **A** Groundwater relative to Leeton rainfall and LMWL and **B** Enlargement for Tubbo site between Sep-98 and Feb-99.

Table 7.6: Slope and deuterium excess for Australian rainfall isotope monitoring stations.

Location	LMWL slope	LMWL ^{2}H intercept	No. of data points
Brisbane *	7.76	12.89	227
Alice Springs *	7.56	9.3	75
Melbourne *	7.44	8.48	224
Adelaide *	8.08	12.17	89
Ballimore [a]	7.76	15.56	4 (2 from Gunnedah)
Gunnedah [b]	7.9	16	2?
Gunnedah [c]	8.4	15.99	17
Leeton	7.3	11.3	10

* calculated from International Atomic Energy Agency data by Scofield 1998
[a]Schofield 1998
[b]Lavitt 1999
[c]Timms & Acworth 2002

The LMWL established may be compared with coastal monitoring stations and a continental station, Alice Springs (Table 7.6). Leeton LMWL has a lower slope and a deuterium excess that lies between the values for Melbourne and Brisbane, the two closest stations. These differences are to be expected, but do suggest that the Leeton rainfall samples are likely to be representative of the long term average for the continental Riverina plain. Furthermore, the sampling period was one of slightly above average rainfall and was of over 12 months duration, capturing both summer and winter rainfall events.

Isotopic signature of groundwater

With respect to this LMWL, the Murrumbidgee River, B1 (Sep-98 to Feb-99) and B4 (Sep-98) were enriched in δ^{18}O, while most groundwater samples were depleted, particularly with respect to δ^2H. Groundwater samples were located on a line of slope 3.4 (Figure 7.9A), with no sample more enriched than the mean rainfall δ^2H and δ^{18}O.

The similarity of deep aquifer and aquitard isotopic signatures and the increasing depletion of B4 groundwater indicates thorough mixing within the alluvial system (Figure 7.9B). Prior to pumping, it is dominated by isotopically enriched water, related either to the shallow aquifer or the river. At least two unknown groundwater compositions appear to be contributing to mixing (unidentified end-member, UEM), an older more depleted groundwater between Sep-98 and Oct-98, and a younger more enriched groundwater after Oct-98. The most significant isotopic changes occur in the first month of groundwater pumping.

The deep aquifer, (B6) and silt unit, (B5) show very small shifts towards isotopically enriched waters between Sep-98 and Oct-98. Since the changes are similar to the detection

Table 7.7: Stable Isotope results for Tubbo site, Sep-98, Oct-98 and Feb-99.

Site	δ^2H $\pm$1 ‰			δ^{18}O $\pm$ 0.2 ‰			δ^{13}C $\pm$ 0.2 ‰	
	Sep-98	Oct-98	Feb-99	Sep-98	Oct-98	Feb-99	Oct-98	Feb-99
River	–	-41	–	–	-6.2	–	-12.0	–
Channel	–	-45.9	–	–	-6.86	–	-15.4	–
B1	-42	-44	-42	-6.47	-6.5	-6.5	-16.9	-17.1
B2	–	–	-48	–	–	-7.4	-15.6	-17.3
B3	-46	-47	-45	-7.43	-7.3	-7.5	-17.9	-17.8
B4	-45.2	-47	-45	-6.5	-7.4	-7.4	-18.0	-16.7
B5	-46.5	-45	-47	-7.05	-7.0	-7.1	-17.9	-17.3
B6	-45.2	-45	-44	-7.28	-7.1	-7.2	-15.7	-17.2
59113	–	–	–	–	–	–	-16.8	–
30284	-44.5	-46	-47	-6.96	-7.0	-7.0	-15.5	-14.2
30282	-44.8	-45	-45	-7.26	-7.3	-7.2	-18.0	-15.6

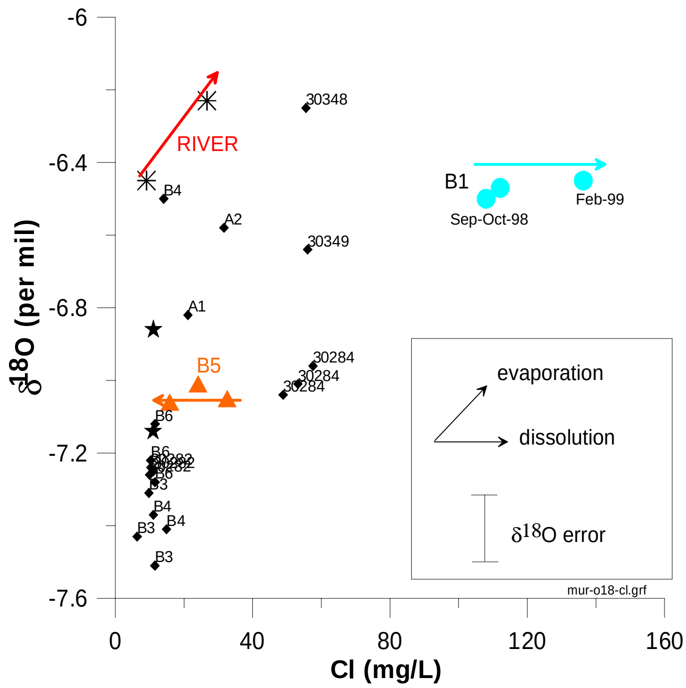


Figure 7.10: Oxygen-18 versus chloride for Murrumbidgee River and groundwaters at Gundaline and Tubbo sites, Sep-98, Oct-98 and Feb-99.

limit, it is uncertain whether this change represents a leakage contribution from shallow groundwater and if so, suggests the proportion to be very small relative to storage.

The relationship between δ^{18}O and chloride in Figure 7.10 may be used to differentiate the origin of chloride in groundwater. The positive correlation between δ^{18}O and chloride for Murrumbidgee River water reflects isotopic enrichment due to evaporation. By contrast, for shallow piezometer B1, increasing chloride is not associated with isotopic enrichment but is the result of dissolution. This result suggests that saline porewater that mixes with the shallow aquifer water is sourced from within the upper silt unit where porewater has a similar isotopic signature, rather than from surface irrigation waters which have a distinct isotopic signature (in this case isotopically depleted, since Tubbo channel water is deep bore water, rather than isotopically enriched river water).

Significant variations were evident for the δ^{13}C data (Table 7.7). Waters relatively enriched in δ^{13}C, may be attributed to exchange with atmospheric CO_2 and dissolution of carbonate.

After 1 month of pumping, groundwater δ^{13}C ranged from -15.5 to -18.0 ‰. The most enriched groundwaters were from piezometers B2, B6 and 30284. By Feb-99, the range of δ^{13}C values at the nested piezometer site had narrowed, with a trend towards more depleted water (Table 7.7). However, ground-water had become more enriched at 30284 and 30282, 1.8 km from the pumping bore, possibly due to lateral flow drawing from the river.

The relatively depleted value for bore 59113, which draws water from deeper in the same aquifer as piezometer B6 indicates that the aquifer is not thoroughly mixed. This may account for the trend towards more depleted δ^{13}C between Oct-98 and Feb-99, suggesting that close to the bore, some upwards component of flow may occur.

Due to the lack of valid stable isotope data for the early stages of pumping, it is only possible to compare data from Oct-98 and Feb-99. Groundwater at B4 appears to mix with a younger water, more enriched in stable isotopes.

7.1.5 Radioisotopes

Radioisotopes, that are naturally decaying in abundance, such as tritium (^{3}H) and carbon-14, ^{14}C, also applied in this study, provide insights into residence time.

Radiocarbon

Radiocarbon dating of groundwater has become a basic tool in hydrogeology since Libby discovered radiocarbon in atmospheric CO_2 in 1946, and the first application of carbon-

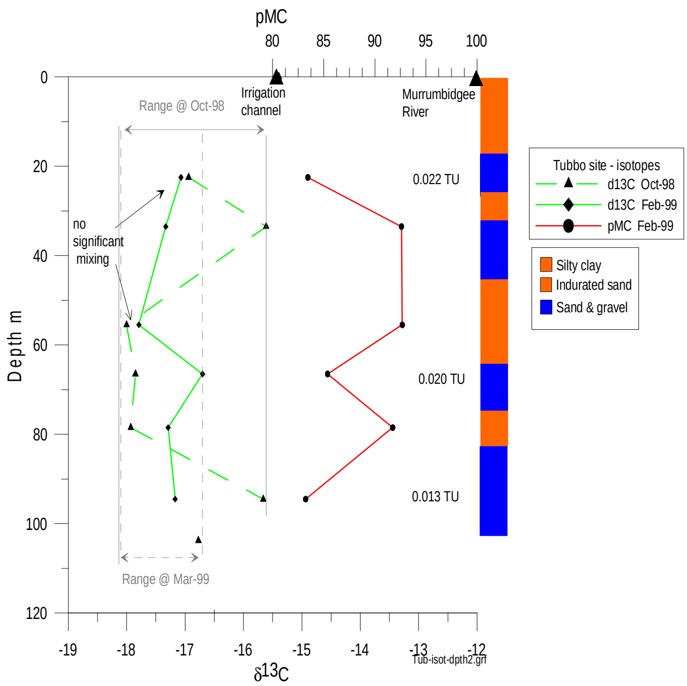


Figure 7.11: Variability of δ^{13}C, ^{14}C and tritium isotopes with depth, Tubbo site, Oct-98 and Feb-99.

14 dating to groundwater in the late 1950s (Sudicky & Frind 1981). An introductory background to ^{14}C dating and standards is found in Chapter 4.3.1.

Higher pMC (of dissolved inorganic carbon) was observed in the clay aquitards (B2, B4, B5), relative to the aquifers (Figure 7.11, Table 7.8). Apparent ages are presented, assuming no mixing and a ^{14}C half life of 5 730 years. Pearson corrected ages are also given along with the assumptions involved in this simplified analytical correction for dilution of ^{14}C. These ages should be regarded as tentative until sufficient hydrochemical and isotopic data from the Calivil aquifer is available for hydrochemical modelling.

Groundwater in aquitards therefore appears to be younger than aquifer water, a finding which conflicts with the fact that residence times are greater in low permeability material. It is possible that these results are effected by addition of dead (or old) carbon by matrix interactions, or that loss of dead carbon may occur within aquitards. Matrix interaction is probably most important within the aquitards which are also characterised by the lowest δ^{13}C values observed in the system. Processes which may account for this unexpected result are to be discussed further later in the chapter.

Table 7.8: Radioisotopes at Tubbo site, Feb-99, including apparent and Pearson corrected ages.

Site	TU [a]	±TU	pMC [b]	Apparent age (years)	Pearson corrected age [c] (years)
B1	0.022	0.017	83.46	1500	1060
B2	–	–	92.62	630	320
B3 [d]	0.020	0.016	92.70	630	530
B4	–	–	85.39	1310	690
B5	–	–	91.76	710	380
B6	BDL	0.017	83.24	1520	1130
30284	0.137	0.016	86.42	1210	30
30282	–	–	87.60	1090	0

[a] TU = Tritium unit, BDL = below detection limit
[b] pMC is percent modern carbon
[c] Pearson correction made using actual δ^{13}C data and assuming matrix carbonate δ^{13}C = 0 ‰ and soil CO_2 of -18 ‰
[d] piezometer samples middle aquifer due to leakage

Tritium

At the Tubbo site near the irrigation bore, very low tritium activity, marginally above detection (0.02 ± 0.017 TU), was found in the shallow and middle aquifer. No tritium was detected in the Calivil aquifer. However in the shallow aquifer 1.77 km away, 0.137 TU was detected.

7.2 Accounting for hydrochemical variability

As noted in the literature, hydrochemical variability can be significant in both a spatial and temporal context. However, prior to evaluating natural or pumping induced variability, it is necessary to identify and either correct or eliminate data which has been affected by the drilling and sampling process.

7.2.1 Artifacts of sampling and well installation trauma

Despite some precautions (see Chapter 4.1.2 & 4.4), a number of samples were affected by well installation trauma due to drilling fluid contamination. The clayey sand of B2 suffered the greatest contamination (3× TDS), with trace contamination in the Lower silt unit B5. Contamination resulted in excess EC and TDS which was primarily due to elevated Cl, HCO_3, SO_4, Na, Sr and Mn.

Decreasing concentrations of these species are evident in Figure 7.7, enabling those samples which are contaminated to be identified, and marked with an '×' on this and all subsequent figures. Contamination persisted in B2 for 5 months (4 of 6 samples) and in B5 for about 2 months (2 of 6 samples). Also, the first sample of B4, in Sep-98, exhibited minor contamination effects. Stable concentrations of contaminate species at levels equivalent to background (eg. with control bores 30282 and 30284) is interpreted as drilling fluid contaminates being adequately flushed from the piezometer intake zone. Increased concentrations thereafter, may be attributed to processes other than well installation trauma.

Variability due to bore casing leakage

Hydrochemical sampling assumes that groundwater pumped is representative of the aquifer (or aquitard) at the depth of the intake screen. However, this may not be the case if leakage has occurred via corroded casings, or via the annulus due to gravel packing or faulty bentonite seals. Anomalies, such as high Fe concentrations in steel bores, rapid decreases in groundwater EC during purging, higher than expected temperatures and spurious groundwater levels may point to possible leakage.

The presence of iron filings and discoloured groundwater strongly suggests that corroded casings may be enhancing vertical leakage at at least 2 sites (Table 7.9). Iron concentrations of 19.2 and 66 mg/L were found at Tubbo (30282) and South DP (30342) respectively. Iron levels above 10 mg/L are probably indicative of corrosion, since Fe levels up to 6.6 mg/L detected in piezometers constructed of PVC suggests such concentrations may be natural.

Variability of groundwater EC during the first few minutes of purging provides another indicator of possible bore casing or annulus leakage. The rapid EC decrease (by a half) observed in 30342 (Figure 7.12) may be attributed to leaked saline water filling the gravel pack, which is rapidly flushed by fresher aquifer water as the capture zone of the pump increases. This interpretation is consistent with the work of Cosler (1997), and Schmidt (1977).

The rapid decrease in EC observed in 30342 is distinct to a delayed and small increase in 30489-1 which is probably due to leakage through the aquifer-aquitard formation beyond the gravel pack of the observation bore. It is significant that 30342 was the only shallow bore to exhibit a rapid EC decrease, which together with high Fe concentrations may be considered very strong indication of casing leakage. However, the absence of an EC trend during purging does not disprove leakage where there is little stratification of EC with depth.

At the Caringal property, one of two irrigation bores (59202) sampled was gravel

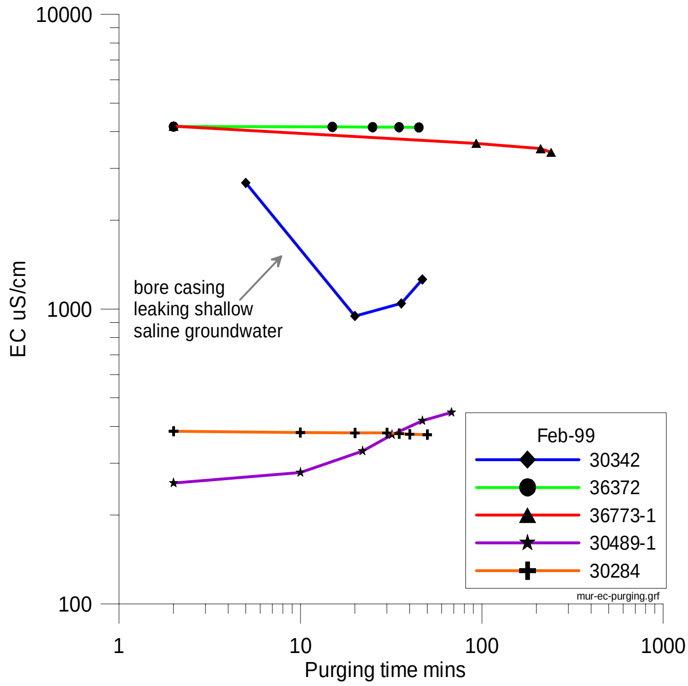


Figure 7.12: Groundwater EC during bore purging prior to sampling, February, 1999.

packed to the surface during construction in 1985 (J. Lot, pers. com.). TDS increased from 113–130 mg/L between 1985 and 1999. This increase of 10% is probably not significant, and may be attributed to relatively fresh shallow groundwater at this location. However, the effects of leakage through gravel packs in areas with saline groundwater may be more significant.

Individual bores may exhibit a unique pattern of variability during purging (Keith et al. 1983), which varies depending screen characteristics, pumping rate, history of pumping, seasonal changes and interconnections in stratified aquifers. The practice of purging a bore prior to sampling, either until field parameters have stablised or 3 bore volumes have been removed, by no means guarantees a 'representative sample'. Careful observations during the first few minutes of this process, however, may reveal the possibility of casing/annulus leakage.

Temperature depth anomalies were not a reliable indicator of casing leakage. Leakage of shallower water should increase groundwater temperature, but it is also possible that steel casings above ground may warm the water column, particularly in summer. This may account for the $+4\,^0$C anomaly in 30284, a bore which was probably not leaking. A temperature anomaly of $+1.5\,^0$C was observed in 30282, while no anomoly was observed

Table 7.9: Indicators of bore casing corrosion and leakage, February, 1999.

Site	Bore	Installation date	Casing material	Fe (mg/L)	Comments
DP south	30342	1-Nov-1972	Steel	66.0	Iron filings
Tubbo	30282	1-May-1972	Steel	1.8 - 19.2	Maximum iron after major leakage event, associated temp & O2 trends
	30284	1-May-1972	Steel	14.2	
RiceMill	36372-1	1-Mar-1980	PVC	BDL	All in separate holes
CIA pump	36774-1	1-May-1988	PVC	BDL	
	36774-2		PVC	6.6	
	36773-1	1-May-1988	PVC	BDL	Separate holes ??
Gundaline	A1 & A2	Jun-1998	PVC	4.2	Newly installed, maximum natural concentration ?
	30348/49	1-Dec-1972	PVC		

in bore 30342, which was likely to be leaking.

Anomalous groundwater levels in 30282 provide further evidence of bore casing leakage (see Chapter 6.1.3). However the absence of water level changes does not disprove casing leakage (van der Kamp & Keller 1993) and borehole leakage may also occur around bentonite seals and gravel packed annulus.

7.2.2 Short term trends

Once extraneous factors contributing to hydrochemical variation have been identified, changes that were observed during the the irrigation season may be considered in terms of enhanced vertical and lateral flow. Although the site was not unfortunately, sampled in Sep-99, prior to the next irrigation season, it is unlikely that groundwater returned to pre-irrigation composition, given the lack of hydrologic equilibrium. Most of the changes occurred within the first couple of months of pumping for irrigation.

Short term hydrochemical changes are evident in the Piper diagram of Figure 7.13. Most notable are the changes in the shallow aquifer (B1). Minor changes in the Lower Shepparton aquifer are unfortunately, largely masked by drilling fluid contamination. Although the early changes in the middle aquifer (B4) can be dismissed due to drilling fluid contamination, this aquifer water is of similar composition as the Calivil aquifer (B6), with later changes toward the composition of either river water or shallow groundwater.

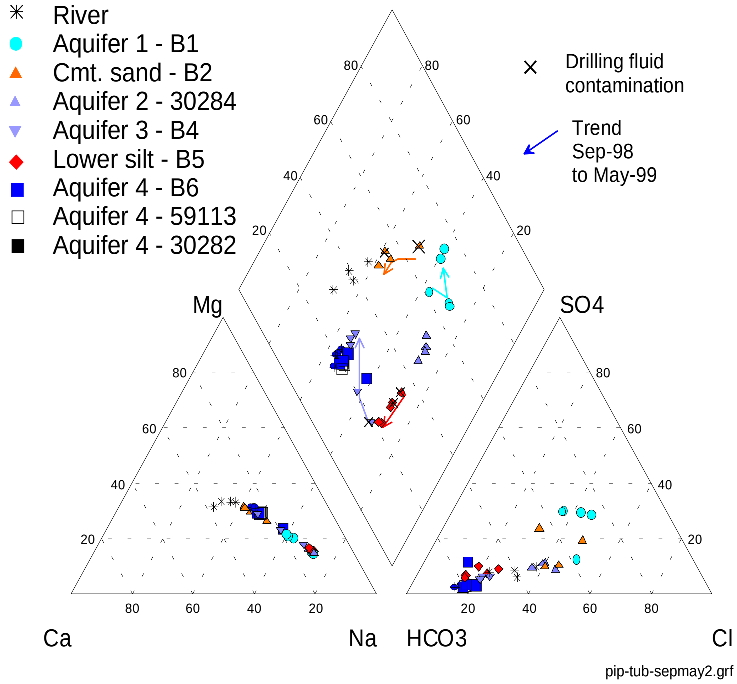


Figure 7.13: Piper diagram of short term hydrochemical changes at Tubbo site, Sep-98 to May-99.

Enhanced vertical flow near a pumping bore

A number of chemical species suggest enhanced vertical flow from the shallow to middle aquifer over the irrigation season. Decreasing concentration with increasing depth suggests downward migration of species for which the sole or primary source is near the surface. Such a depth profile is evident for the following parameters: EC, DO, Na, Ca, Mg, K, Sr, NO_3-N, SO_4 and Cl (see Figures 7.2, 7.2 and 7.8). Higher pH and HCO_3 are also present in the shallow aquifer. For example, since Cl decreases with depth, the primary source of Cl appears to be in the unsaturated clay near the surface, with a very minor source within the Lower silt unit.

Enhanced vertical flow would result in increased concentration of the conservative species over the irrigation season (Figure 7.7). However the magnitude, timing and duration of changes depends on the composition of leaking groundwater and the proximity to hydraulic changes due to pumping. For example in aquifer 1 (B1) Cl increased steadily between Oct-98 and May-99 from 108 to 160 mg/L (+48.1%). However in aquifer 3 (B4) Cl increased from 11.1 to 15.4 mg/L (+38.7%) between Oct-98 and Dec-98 followed by a decline. A small Cl increase (+12.9%) occurred over this time in aquifer 4.

It is evident that significant hydrochemical changes may occur over a scale of weeks to months. Such results concur with (Schmidt 1977) who questioned the commonly accepted adage that 'groundwater quality changes slowly'. Increased vertical flow within the cone of depression of a pumping well was a factor which Ayers & Gosselin (1995) used to explain increased nitrate contaminates in a alluvial aquifer system in Nebraska.

Enhanced lateral flow within aquifers

In aquifer 4, the steady decrease in Cl concentrations away from the pumping bore (30282), and the rapid dilution of the Cl peak (Figure 7.7), indicates that enhanced lateral flow occurs. Freshening of groundwater as it is drawn from the direction of the Murrumbidgee River is to be expected. However such an interpretation assumes that Cl input from vertical leakage is constant between the nested piezometers (B6) and the observation bores (30282), a distance of about 1.8 km.

A proportion of the changes occurring within aquifers may be attributed to hydrochemical variability within the aquifers. Such spatial variability is caused by geological heterogeneity and differences in recharge (Pettyjohn 1976; Pettyjohn 1982) who introduced the concept of cyclic changes in groundwater quality in the absence of anthropogenic factors. Constant 'background' concentrations are in some cases, an invalid assumption, particularly in environments where unsaturated zone storage and macropore flow occurs.

7.2.3 Medium term trends

Historical data was difficult to interpret due to infrequent sampling and lack of information regarding methodologies and instrumentation. Very little historic data is available, partly due to a lack of regular, representative testing, the difficulty of obtaining data from a variety of electronic and print formats stored in central and regional offices and the exclusion of original data in reports. The quality of groundwater data may vary significantly depending on how thoroughly bores are purged prior to sampling, and to accurate calibration and temperature correction of field meters. Despite these technical difficulties, the magnitude of trends was in many cases clearly greater than uncertainty in data quality.

The relatively high salinity of groundwater between 49–60 m may be due to further downwards movement of the salinity bulge identified by van der Lelij et al. (1987). As noted in Chapter 2.7.4, this salinity buldge which was originally at a depth of 16–19 m had migrated downwards since irrigation to a depth of 30 m. Since the original EC data from this analysis is not available, it cannot be known whether higher salinity at 49–60 m is a result of the salinity bulge migrating downwards over the past 20 years.

Significant changes occurred for the shallowest bores at 2 sites 30342 and 36372-1. The rate of change in 36372-1 appears to be relatively constant, between 360–380 μS/cm/year over the last decade. Conductivity in 36372-1 has increased by a factor of 14 over a decade. This is well above potential salinity increase estimated by Williams (1987).

The increasing trends at the majority of sites, suggest that leakage is likely to be widespread, if not regional. At one site however, the trend is at least partially enhanced by localised leakage through corroded bore casing (eg. Darlington Point south site).

7.3 Geochemical processes

7.3.1 Leakage from clayey aquitards

Trace elements such as Li provide sensitive tracers of leakage of porewater which has resided within clay aquitards.

Release of lithium from clays during infiltration usually results in increased lithium relative to Cl and Na. Micas and amphiboles are also a Li source (Reimann & Caret 1998). The Li/Cl ratio was found by Edmunds (1994) to be proportional to residence time with up to 10^5 years required to reach equilibrium for a particular formation. It is therefore to be expected that groundwater from shallow clayey sediments (12986 and 12985) contained the greatest Li concentrations, 0.064 and 0.046 mg/L respectively.

Li concentrations above detection were not commonly found until May-99, possibly indicating leakage from clays during the 1998/98 irrigation season. Li decreased with depth at the Tubbo site, with an unexpectedly high Li level in B6. High Li in 30282, particularly prior pumping (0.04 mg/l in Oct-98) is consistent with other evidence of episodic leakage from overlying clays.

7.3.2 Mixing and dilution

A mixing diagram was constructed using chloride concentrations of end members, the Murrumbidgee River, and the shallow saline groundwater (12986) (Figure 7.14). At high mixing ratios, HCO_3 is non-linear indicating processes other than mixing controlling concentration. There are also some non-linear SO_4 samples for high mixing ratios. Gain of Na in fresh water samples is evident, indicating ion-exchange is an important process which will be discussed further.

Mixing of groundwater from different layers in the system was quantified with mass balance geochemical modelling, which is to be presented later in the chapter.

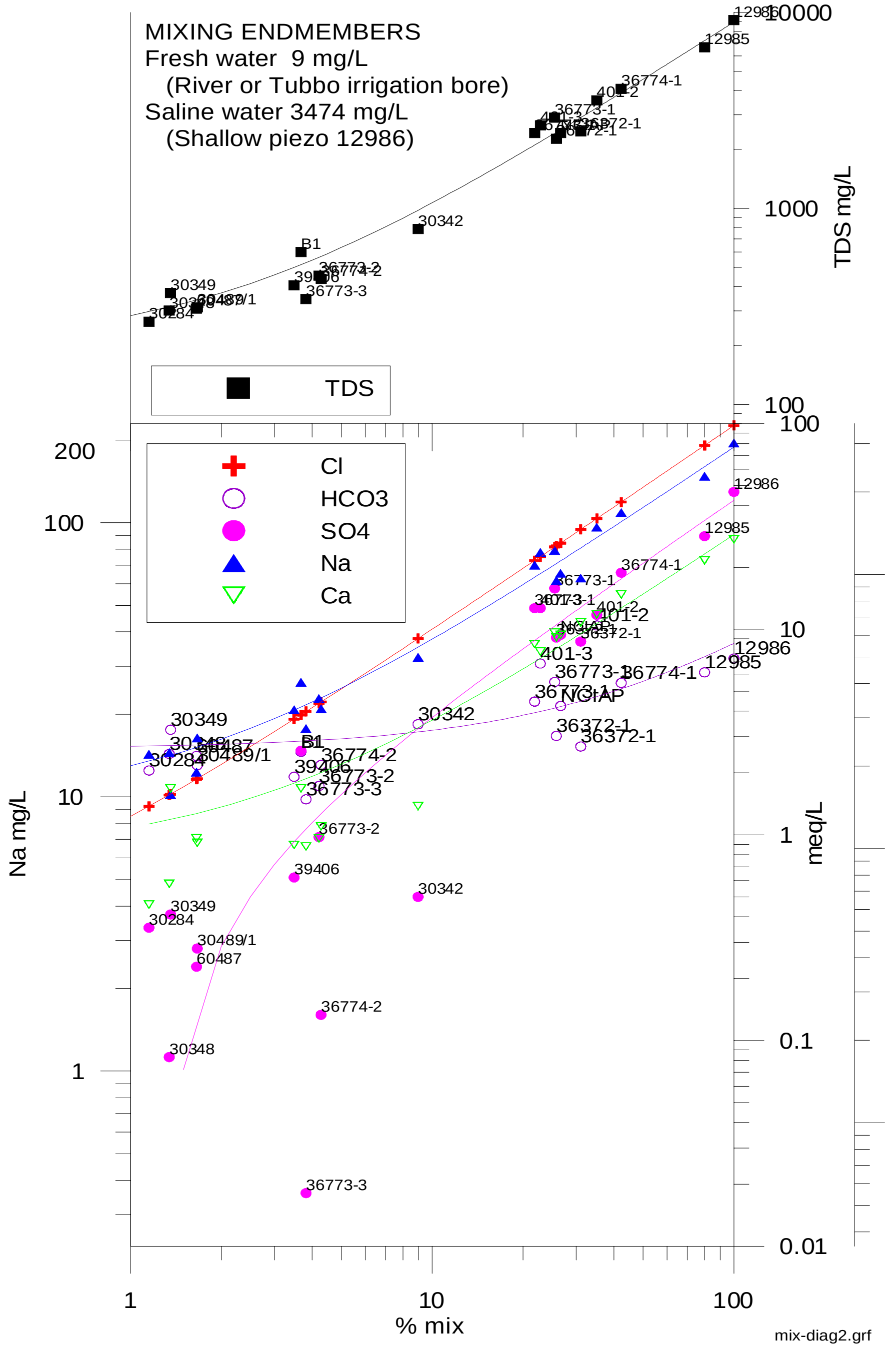


Figure 7.14: Mixing diagram for low chloride river and deep aquifer water with high chloride shallow groundwater.

Mixing with palaeowaters

Isotopic evidence which indicates interaquifer mixing, will now be considered together with data that suggest deep groundwater to be dominated by palaeorecharge. Variable stable isotope data for the Tubbo site during the irrigation season has been described, including groundwater becoming increasing enriched in δ^2H data and convergence of δ^{13}C data (Figure 7.11).

Yet despite hydrochemical and isotopic evidence for mixing in the middle and deep aquifers, groundwater at this depth remains dominated by palaeorecharge. All groundwaters were depleted in δ^{18}O and δ^2H relative to average modern rainfall (Figure 7.9), which indicates recharge under wetter and cooler climatic conditions (Clark & Fritz 1997).

Although the lengthy residence time of deep groundwater, appears to conflict with evidence for recent vertical leakage, the evidence is not in fact contradictory. Matthes et al. (1976) pointed out practical problems of groundwater age dating in relation to recent contaminated recharge. For example, a 54 000 year old water mixed with 1% modern recharge would reduce the apparent groundwater age by 33% to 36 100 years. It is possible then, that groundwater with an apparent age of thousands of years at Tubbo, could be a mixture containing modern recharge.

The mixing interpretation is supported by tritium, which was marginally above detection limit, in the middle aquifer. If tritium input of 4.3 TU is assumed, appropriate for modern Australian rainfall (Calf 1987), such a low tritium activity could be interpreted as unmixed groundwater in which tritium concentration has decayed since recharge about 100 years ago (over 7 half lives of 12.43 years). Alternatively, such a low level of tritium could be interpreted as a mixture $\sim$0.5% rainwater, that has leaked via rapid flow pathways, with palaeowater that contains no tritium (eg. groundwater from the Calivil aquifer). The later explanation is consistent with other hydrochemical evidence for mixing.

A more significant tritium activity of 0.137 TU at the DLWC bore 30284 is even more convincing of vertical mixing. In this case, groundwater could be either unmixed recharge from about 60 years ago, or a mix of $\sim$3% modern recharge with older groundwater. The second possibility is consistent with geophysical evidence of a buried aquitard window at this site (Chapter 5.1.2), and the excessive transit time required for modern recharge to flow laterally from the river. At least 500 years is required for river recharge to reach the Tubbo site, at a linear velocity of 12 m/year estimated on the basis of hydraulic data by (Drury et al. 1984). Over this period of lateral flow all tritium sourced from the river would have decayed below the detection limit, so the detection of tritium is strong evidence for a component of vertical recharge at the site.

7.3.3 Dissolution and precipitation

An indication of whether groundwater is in equilibrium with solid mineral phases and therefore whether dissolution or precipitation is likely to occur is given by saturation index (SI). For example, SI is defined for calcite as follows:

$$SI = log\frac{IAP}{K_{sp}} = log\frac{[Ca^{2+}][SO_4^{2-}]}{K_{gypsum}} \tag{7.3}$$

Where IAP is ion activity product of the activites of species in solution and K_{sp} is solubility product.

At equilibrium, saturated solutions are those defined by an SI >0 where precipitation may potentially occur. Undersaturated solutions are characterised by SI <0.

SI may be calculated manually, or with a speciation computer program such as WATEQ4F (Ball & Nordstrom 1991) which is able to account for full ionic strength and K_{sp} temperature correction. An error of ±0.1 SI is to be expected due to combination of laboratory error, sample contamination and inaccuracies in the thermodynamic data set.

Anhydrite and gypsum were undersaturated in all samples, and calcite and dolomite undersaturated in most samples (Figure 7.15). Precipitation of calcite and dolomite may potentially occur where oversaturated (12985 and 12986). Saturation indices of approximately zero were calculated for 36774-1, and 36773-1. Siderite ($FeCO_3$) was oversatured in a number of samples, particularly those with high iron. Slight supersaturation of siderite is a common situation in groundwater (Appelo & Postma 1996).

There was no apparent trend of SI with depth, though aquifers between approximately 95 to 120 m depth exhibited similar SI to Murrumbidgee River water.

The dissolution of NaCl is indicated by the 1:1 ratio of NaCl at higher Na and Cl concentrations, and also by the Na/Cl versus EC for higher salinity waters (Figure 7.16).

7.3.4 Ion exchange

Ionic ratios offer clues as to geochemical processes. Ion exchange involves the replacement of one chemical for another at the solid surface. Ion exchange is particularly important in dynamic environments where the flow of groundwater of different composition triggers ion exchange, which acts as a temporary buffer in non-steady state situations such as a moving

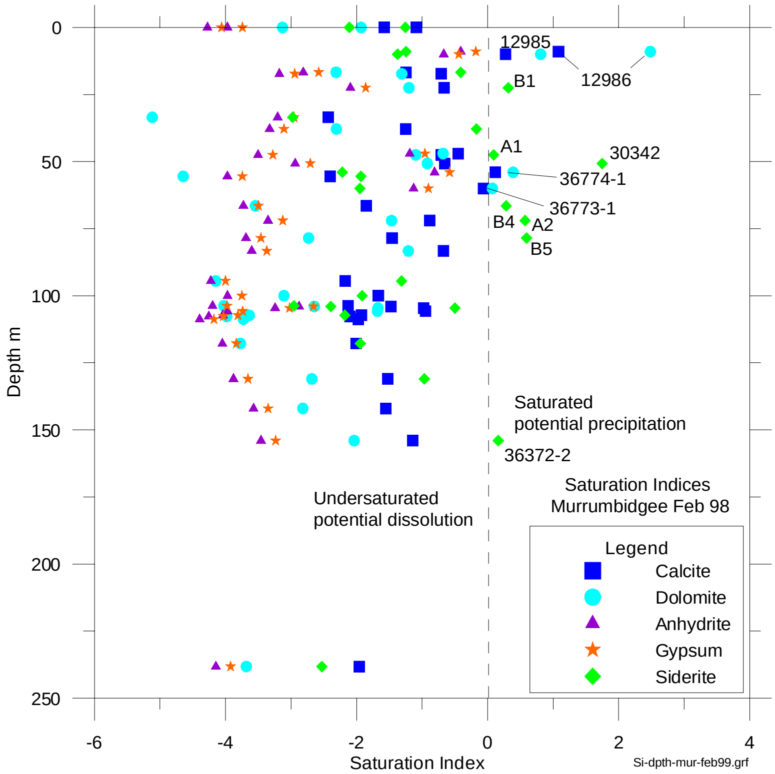


Figure 7.15: Saturation indices for groundwater and surface waters, Feb-98.

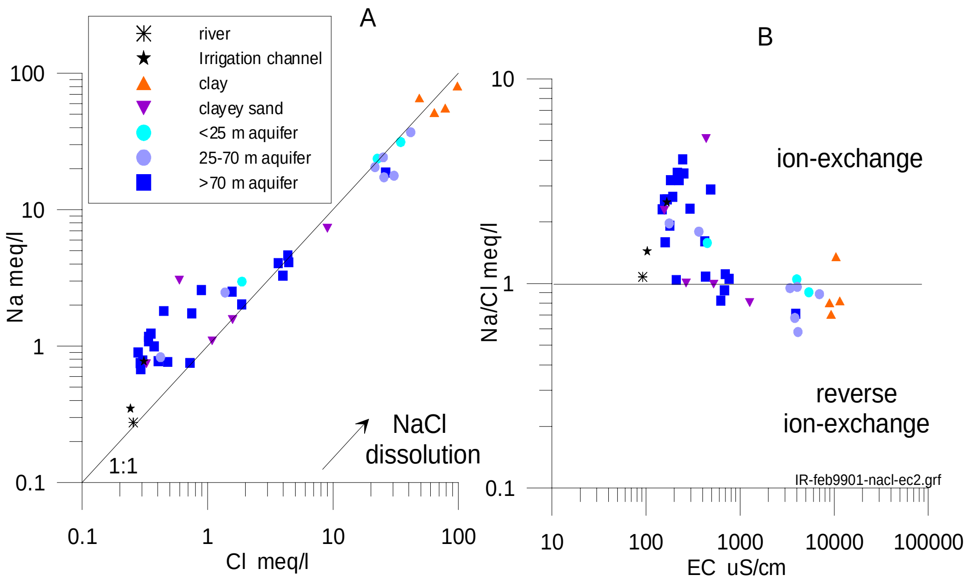


Figure 7.16: Ionic ratio bivariate plots for ground and surface waters, Feb-99 & Feb-01 (Additional ionic ratio bivariate plots can found on the CD-ROM).

saline interface (Appelo & Postma 1996). Water with similar ionic ratios, normally calculated in terms of milliequivalents, have generally evolved by common hydrogeochemical reactions. Ion exchange reactions occur rapidly, and where flow velocities are less than about 100 m/year, equilibrium can be assumed (Appelo & Postma 1996).

Reverse ion exchange requires clay with exchangeable Ca or Mg (usually montmorillonite) and a flowing water to have a higher concentration of Na relative to the clay matrix to drive a reaction as the system equilibrates. Relationships between dissolved species (see Na/Cl versus EC Figure 7.16B) indicate that reverse ion exchange is an important process in the evolution of all shallow clay groundwaters and some shallow sand groundwaters which are likely to be a result of mixing.

Reverse ion exchange is typically observed with seawater intrusion or brine contamination. At the saline intrusion front, Na is exchanged for Ca. However once calcium in the clay has been removed, the process cannot continue further - thus $CaCl_2$ type water occurs at the leading edge of contamination (Hounslow 1995). This process is clearly seen in the evolution of aquifer water at the Rice Mill site between 1980 and 2001 (Figure 7.4 A). Between 1980 and 1995 the proportion of Ca relative to Na increased as the water became increasingly saline. However, between 1995 and 2001, the direction of this trend had changed with the relative increase of Na. This change presumably represents the stage of Ca removal from exchange sites within the aquifer and at the aquifer-aquitard interface. Similar, but less distinct trend changes were observed at the Darlington Point south and east sites sites (Figure 7.4 C-D).

7.4 Mass balance modelling

7.4.1 Quantifying mixing and matrix interaction

A hydrogeochemical model facilitates calculations in complex problems, and provides the means to test concepts and a suite of reactions for a given environmental problem (Appelo 1996). Geochemical modelling techniques were first developed by Garrels and Mackenzie (1967) at a time when computer technology availablility made the many complex calculations feasible (Chapelle 1993). The development of geochemical models since then are reviewed extensively by Plummer (1984), Parkhurst & Plummer (**?**), and Chapelle (1993), to which the reader is referred for a more comprehensive introduction to this subject. It is important to note here, that inverse modelling in this study uses real available groundwater chemical data along a flow path to determine net geochemical processes that satisfy the principle of conservation of mass.

Mass balance model codes

The computer program WATEQ4F (Plummer et al. 1976) was used as the primary speciation tool, as it is particularly suited to low-temperature, low-salinity water where carbonate equilibria are important. Preliminary mass balances were made with NETPATH v2, an interactive Fortran program which computes mass-balance reactions between initial and final waters on a flow path (**?**).

Subsequent models were developed with the PHREEQC code (**?**), which includes the ability to account for uncertainity in chemical and isotopic data. This facility, together with the potential to couple with a flow code for reaction transport modelling means that PHREEQC has been applied to an increasing range of forward and inverse modelling problems.

7.4.2 Assumptions

Appelo & Postma (1996) discuss several limitations of mass balance modeling in identifying possible reactions.

- Mass balance modelling assumes steady state conditions and that there is no temporal variability in water entering the system. Implicit in this assumption is that losses due to advective dispersion are insignificant.
- Flow path is hydrologically feasible.
- Assume homogeneous reaction between points of analysis - no matrix heterogeneity.
- No unique solution - results are not proof that reactions actually occurred.
- Thermodynamic constraints are not included and models may not be kinetically consistent. Implausible reactions must be eliminated by checking saturation indices.
- Only reactions that change net water composition are considered.
- System is assumed to be invarient in terms of temperature and pressure.

For modelling in this project, it has also been assumed that redox reactions do not contribute significantly to results as all the waters sampled are oxidised and contained free O_2. Trace elements were not included in the mass balance models, as there is often uncertainty in concentration levels due to sorbtion on mineral surfaces (**?**).

The shallow aquifer was assumed to be in steady state, by choosing periods when changes were relatively constant (eg Dec-98 to Feb-99). This approach, it may be argued, is a gross simplification of reality since the Cl increase in B1 was actually episodic (+1.2 mg/L beween Oct-98 and Dec-98, +27 mg/L between Dec-98 and Feb-99 and +25 mg/L between Feb-99 and May-99). However, in the absence of a more comprehensive dataset with which to implement transient reaction transport modelling, these models serve as a first step in identifying important processes.

7.4.3 Rice Mill site

Detailed analysis of groundwater leakage and mixing at the North Coleambally site (36372) was possible on the basis of a number of chemical analyses between 1980 and 1999. At this site, located to the north of the irrigation area, rising groundwater levels between 1979 and 1996 in both the Shepparton and Calivil coincided with increasing seasonal drawdown. By 1997, drawdown in the Calivil aquifer was so great that seasonal recovery was minimal and groundwater levels in the Shepparton aquifer had declined. The downwards hydraulic gradient between the Shepparton and Calivil aquifer in February, 1999 was 0.052.

Over this period of time, groundwater evolved from Na-HCO_3 to Na-Mg-Ca-HCO_3-Cl type (Figure 7.5), and MH for irrigation increased from 48.8 to 57.3, which is above recommended levels, while the SAR remained low, decreasing from 5.6 to 5.0.

A two component mix was modelled, consisting of 36372-$1_{(Sep-86)}$ and a saline Shepparton clay groundwater (12985). The aim was to produce the groundwater composition observed in 36372-$1_{(Feb-99)}$. Several possible models were developed and the 7 plausible phases consisted of: $CaCO_3$, $CaSO_4$, $CO_{2(gas)}$, KCl, NaCl, Na/Ca and Na/Mg exchange. For the exchange reaction the first species exchanges onto the clay and the second species is released into solution.

Based on available data, representative mass transfer, in mmol/kg of water, is:

$$
\begin{aligned}
&(70\%\, of\ 36372-1_{(Sep-86)}) + (30\%\, of\ 12985) + 6.62 NaCl + 0.01 KCl \\
&\Rightarrow 36372-1_{(Feb-99)} + 1.2 CO_{2(gas)} + 0.88 CaCO_3 \\
&+1.11 Na/Ca_{exchange} + 2.22 Na/Mg_{exchange}
\end{aligned}
\tag{7.4}
$$

Preliminary estimates of Shepparton Clay groundwater in the mix varied from 10 to 38% depending on the composition of the saline end member (12985 or 12986), and the forcing of plausible phases (Timms et al. 2000).

To further constrain these estimates, stable isotope data, and mineralogical data for the Shepparton clay and site specific saline end member was sought. A second phase of drilling in November, 2000 installed three additional piezometers at this site, of which the shallow one was dry. Chemical data from the 12 m deep piezometer 401-2 which was screened in saline clay, was then used as the saline endmember in subsequent modelling.

Table 7.10: Mass balance model development for Rice Mill site, between Initial 36372-1$_{(Sep-86)}$ and Final 36372-1$_{(Feb-99)}$ Aquifer waters

Simulator setup	Simulation ID	Saline porewater	Constraints	Plausible models	Mix % of saline porewater
NETPATH (Timms et al.2000)		12985	Calcite, gypsum, halite, sylvite, $CO_{2(g)}$, Ca/Na$_{ex}$, Mg/Na$_8$	5	10 - 38
NETPATH site specific saline porewater	Rm2br	401-2	Calcite, gypsum, halite, sylvite, $CO_{2(g)}$, Ca/Na$_{ex}$, Mg/Na$_{ex}$	2	73 - 88
PHREEQC uncertainty 10%	rmill11	401-2	Calcite, gypsum, halite, sylvite, $CO_{2(g)}$, Ca/Na$_{ex}$, Mg/Na$_{ex}$	1	86 - 91
PHREEQC uncertainty 5%	rmill13	401-2	Calcite, gypsum, halite, sylvite, $CO_{2(g)}$, Ca/Na$_{ex}$	1	96 - 98
PHREEQC uncertainty 5%	rmill13	401-2	Calcite, halite, sylvite, $CO_{2(g)}$, Ca/Na$_{ex}$, Mg/Na$_{ex}$	1	79 - 82

There was a large difference in the estimated proportion of saline leakage (Table 7.10) using a non-site specific composition (12985) and a groundwater analysis from the site (401-2). This highlights the importance of using site specific saline porewater composition for mixing rather than an average regional value. Utilising an area averaged saline porewater could lead to unacceptable errors in the proportion of mixing.

The advantages of using the new inverse modelling functions of PHREEQC which include uncertainty analysis are also evident. Mixing estimates using identical input parameters and constraints were similar, althought the PHREEQC results were at the upper end of the NETPATH range. However, the minimal model function of PHREEQC generated a single model, for which a range of mixtures which satisfied the prescribed (10%) analytical uncertainty of input data.

Additional nested piezometers are necessary to detail geochemical and mixing processes within the clay dominated segment of the flow path. Calcite, for example, is undersaturated in the Shepparton aquifer, but may have precipitated earlier along the flow path within the Shepparton clay.

7.4.4 Tubbo site

Mass balance modelling of mass transfers indicates that similar chemical reactions occur during mixing of Shepparton clay pore waters with Shepparton aquifer groundwater, to that discussed for the Rice Mill site. Hydrochemical processes were dominated by NaCl dissolution, CO_2 outgassing and reverse ion exchange. The composition of $B1_{(May-99)}$ is developed by a mixing of about 20% of Shepparton clay pore water with 80% of $B1_{(Oct-98)}$. Based on available data, a representative mass balance was:

$$\begin{aligned}(80\%\, of\ B1_{(Oct-98)}) + (20\%\, of\ ShepCL) + 0.38NaCl \\ \Rightarrow B1_{(May-99)} + 5.4CO_{2(g)} + 0.4CaCO_3 \\ +0.7Na/Ca_{ex} + 0.24Na/Mg_{ex} + 0.04K/Na_{ex}\end{aligned} \tag{7.5}$$

Table 7.11: Mass balance model development for Tubbo site, between Initials B-1_{Oct98}, Shepparton Clay porewater and Final $B1_{May99}$.

Simulator setup	Mix % [a]	Calc	Gyp	Hal	Syl	$CO_{2(g)}$	Ca/Na_{ex}	Mg/Na_{ex}	K/Na_{ex}
NETPATH [b]	20.3	-0.48	-	0.42	-0.04	-5.5	-0.71	-0.24	-
PHREEQC [c]	13-23	-0.22	-	0.59	-0.03	-6.12	-0.47	-0.30	-
PHREEQC [d]	13-23	-0.22	-	0.59	-	-6.12	-0.47	-0.30	+0.03
PHREEQC [e]	19-36	-0.22	-	-	-	-5.6	-0.43	-0.23	+0.04
PHREEQC [f]	12–41	-	-	-	-	-4.85	-0.22	- 0.24	+0.04

[a] % Mix of saline porewater. All transfers are in mmol/kg where + indicates dissolution and - indicates precipitation. Ion exchange species listed in order of appearance in solution.
[b] Timms et al. 2000, Model: Tub-4
[c] Model: B1IAH1
[d] Model: B1-2
[e] 5% uncertainty, Model: B1-3
[f] 10% uncertainty, Model: B1-4

The proportion of Shepparton clay groundwater in the mix varied from 12 to 41% depending on the forcing of certain phases. The sensitivity of these results to pH was tested for PHREEQC model B1-2, since the pH of Shepparton clay porewater was not able to be measured and was arbitarily set at pH 7. Balances ranges of 0.1 and 0.5 pH units were modelled (Model ID's B1-5 and B1-6) with no difference in resulting mass balances.

Mass balance modelling with silicate minerals included as phases was also attempted. Silicate minerals known to occur in the profile (Timms & Acworth 2002b) included quartz, biotite, illite and smectite. However, no models resulted in reaction of silicate minerals such as biotite, (with uncertainty limits varied between 2-5%). It appears that over the short term, silicate mineral weathering is not an important determinate of hydrochemical composition.

7.5 Summary

- Salinisation of the aquifers to a depth of 50 m was observed at all 4 sites away from the river over the medium term (since mid 1980s). Minor freshening occurred at the 2 sites near the river, but no corresponding trends were observed in the Calivil aquifer over the medium term. Localised NO_3 contamination of the shallow aquifer was detected.

- Hydrochemical changes occurred in all aquifers at the Tubbo site over the 1998-99 irrigation season. A large, but delayed TDS increase was observed in the Upper Shepparton aquifer which persisted beyond the end of the irrigation season. Soon after pumping began, a small pulse of saline groundwater was observed in the Lower Shepparton aquifer and a very small pulse of saline groundwater was detected in the Calivil aquifer though this latter pulse dissipated within weeks due to lateral influx of fresh groundwater.

- Mixing of the Lower Shepparton and Calivil aquifers was confirmed by the convergence of stable isotope data, although radioisotopes, and the depletion of $\delta^{18}O$ and $\delta^{2}H$, relative to the LMWL indicated that palaeorecharge dominated the deep aquifers.

- According to mass balance hydrogeochemical models, hydrochemical trends in the shallow aquifer, over the long term at Rice Mill site could be accounted for by mixing aquifer water with 80% saline porewater, and at the Tubbo site during the 1998-1999 irrigation season, by mixing with 12–41% of saline porewater. Hydrochemical and isotopic data indicated that this salt originated by dissolution of NaCl stored in the upper silt unit, rather than by recycling of evaporated irrigation water.

Chapter 8

SYNTHESIS

It is now appropriate to integrate hydrogeological, hydrodynamic and hydrogeochemical aspects of the project. This chapter provides a synthesis of important solute transport mechanisms. Leakage through aquitard windows, when the groundwater system is under stress is related to hydrochemical changes.

8.1 Solute transport processes

The physical characteristics of aquitards, especially permeability and spatially continuity, determine which solute transport processes are dominant.

8.1.1 Flux through the vadose zone to the shallow aquifer

It was not within the scope of this thesis to examine flux mechanisms through the vadose zone, although a few comments will be made. Clayey silts within the vadose zone are the most significant source of dissolved solutes to the groundwater system (Timms et al. 2000). Downwards flux of these solutes accounts for increased TDS of the shallow aquifer, over the medium term, and during the irrigation season at Tubbo site (Figure 8.1).

Hydrochemical and isotopic evidence has shown that recharge to the shallow aquifer is saline due to dissolution of NaCl occurs, rather than evaporation (Figure 7.10). Tritium data (Table 7.8) showed that greater vertical leakage occurred below the DLWC bore (30284) compared with nested piezometer (B1). This was possible because of a major aquitard window buried by a veneer of clay at the DLWC bore (Figure 5.4A electrical image).

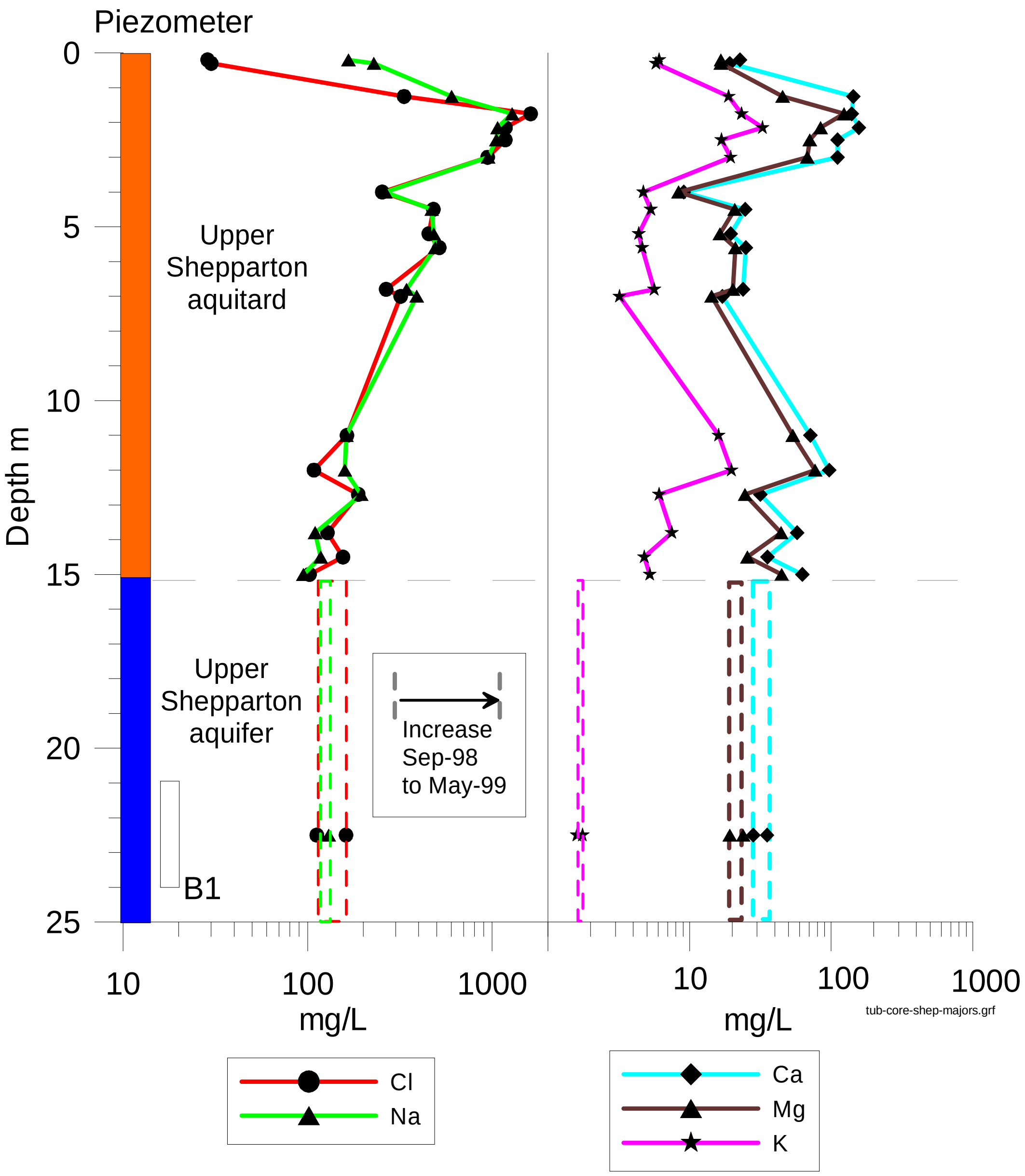


Figure 8.1: Relationship betweeen major ion chemistry of porewater determined from Upper Shepparton aquitard core samples and groundwater from the underlying Upper Shepparton aquifer.

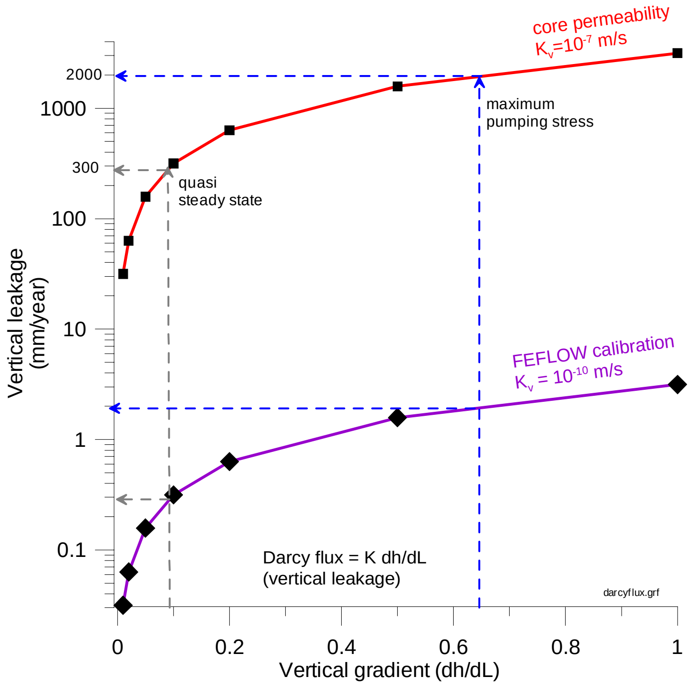


Figure 8.2: Range of Darcy flux for the indurated sand aquitard under natural and induced gradients, based on measured and calibrated model permeability values.

8.1.2 Advective flux between the shallow and middle aquifers

Since the clayey sand forms an effective hydraulic barrier between the shallow and middle aquifers, transport of solutes must occur through, rather than around the aquitard. The Peclet number (Equation 3.9)was to determine the dominant mechanism for solute transport based on measured aquitard hydrualic parameters. Using $\phi_e = 0.09$ (Chapter 5.1.8), L = 6 m, and $D_e = 2{\times}10^{-10}$ m^2/s, Pe was calculated to be >3000. Advective transport thus dominates over diffusive transport.

If this unit is assumed to be laterally extensive, Darcys law may be applied to estimate leakage rate, using observed hydraulic gradients and measured hydraulic conductivity. The relationship between Darcy flux (vertical leakage rate) and K_v and dh/dL, is illustrated in Figure 8.2.

Darcy flux between the Upper and Lower Shepparton aquifers at Tubbo site was then calculated to be 0.3 mm/year prior to pumping and 2 mm/year during the maximum pumping stress of the first seasonal pumping cycle. It should be noted however, that these fluxes occurred 35 m from the irrigation bore, and are not reflective of average values in the total area of pumping influence.

The range of values reflects uncertainty in the permeability of the clayey sand that was derived from core testing (10^{-7} m/s) and model calibration (10^{-10} m/s). Since permeability scale effects generally result in a large modelled K compared with core testing, the 3 order of magnitude difference is attributed to heterogeneity within the indurated clayey sand. The importance of testing permeability on a range of scales, both by laboratory and field testing, and by numerical modelling is evident.

8.1.3 Leakage through aquitard windows between the middle and deep aquifers

Direct mixing of water of the Lower Shepparton and Calivil aquifers takes place via aquitard windows. Minor downwards dispersion probably occurs in the absence of large groundwater fluxes, but this process is greatly enhanced by pumping.

Hydraulic evidence for aquitard windows at this depth was presented in Chapter 6.2, including the observation that drawdown that occurred in the aquifer overlying the Calivil aquifer, prior to drawdown within the intervening aquitard. The timing and magnitude of hydrochemical changes that were observed within each aquifer after pumping was initiated indicate that rapid inter-aquifer flux occurs (Figure 7.7). In constrast, aquifer-aquitard interactions were important over the medium to long term.

8.1.4 Diffusive flux between aquitards and deep aquifers

Diffusive flux is the dominant transfer mechanism of salts between the Calivil aquitard and aquifer. Advective transport does not occur, partly because of the lack of gradient between the interconnected aquifers adjacent to the aquitard. The Peclet number for this 8 m thick aquitard was estimated to be ≤ 2.4, based on K = 10^{-11} m/s, $\phi_e \sim 0.05$, dh/dl ≤ 0.3, and $D_e = 2\times10^{-10}$ m^2/s substituted into Equations 3.9. Since the Pe is significantly less than 5, this confirms that diffusion is in fact dominant.

Although the Calivil aquitard is not a major salt store, the variability of major ions in aquitard porewater indicates diffusion of salts into adjacent aquifers has occurred over the long term. It is possible that a reversal of concentration gradients more recently, has resulted in diffusion from the Calivil aquifer into the edge of the overlying aquitard.

Figure 8.3 shows that peak TDS concentration of 226 mg/L, occurs approximately in the middle of the 7 m thick aquitard and declines steadily to about 16 mg/L at the interface with the overlying aquifer, and about 85 mg/L with the interface of the underlying Calivil aquifer. By comparison, salinity of adjacent aquifer water is higher at 128–131 and 126–143 mg/L TDS, respectively for the overlying and underlying aquifers, sampled between Oct-98 and May-99.

The linear decrease in Ca, Mg, and HCO_3 concentrations (towards the edges of the aquitard) contrasts with both highly variable Cl and Na which is relatively constant with depth. The concentrations of Ca, Mg, and HCO_3 of aquitard porewater are very similar to those measured in water sampled from the piezometer B5, within the aquitard. K was below detection limit in aquitard porewaters, but was of a similar concentration in piezometer B5 within the aquitard, and aquifer waters. However, Na, and Cl concentrations in piezometer water appear to be a mix of aquitard porewater over the 3 m length of screen, which also included a high Cl peak. High Cl porewater at 80.5 m, possibly reflects analysis of a relatively unweathered clay relict.

Diffusive flux of selected species from the middle to the lower edge of the Calivil aquitard was estimated using Fick's first law. An effective diffusion coefficient, D_e of 2×10^{-10} m^2/s was adopted. Ca was selected as the diffusive species, with a concentration from 13.4 to 3.8 mg/L in the centre and edge of the aquitard respectively, a transport distance of 3 m. This relatively steep concentration gradient results in an estimated downwards Ca flux of $20.2\,mg/m^2$ year.

The total flux of Ca, Mg and possibly Cl, would thus be significant, when compared with estimated flux of NaCl from the Geera clay to the Renmark aquifer of 1 g/m^2 year (Hanor 1987), although this estimate may have been underestimated. This is the case because a diffusion coefficient was estimated using a geometric tortuosity factor of 3 m/m resulting in D_p of 7.5×10^{-11} m^2/s. In fact, the complex tortuosity factor appropriate for clay is less than the geometric tortuosity factor, due to the effects of the double-layer decreasing fluidity, and electrostatic interaction (Rowe et al. 1995).

Flux of saline porewater to the Calivil aquifer must have occurred sometime in the past, and is no longer active, because concentrations of these species in adjacent aquifers currently creates a diffusive flux into, rather than out of the aquitard. It is possible that the relatively higher Cl concentrations in aquitard porewater at the lower edge may reflect diffusion back into the aquitard at this interface. Rowe et al. (1995) note that Cl has a relatively high diffusion coefficient in free solution.

The aquitard may not be sufficiently thick to preserve original solute concentrations, particularly since several pore volumes are estimated to have flushed the aquitard since deposition.

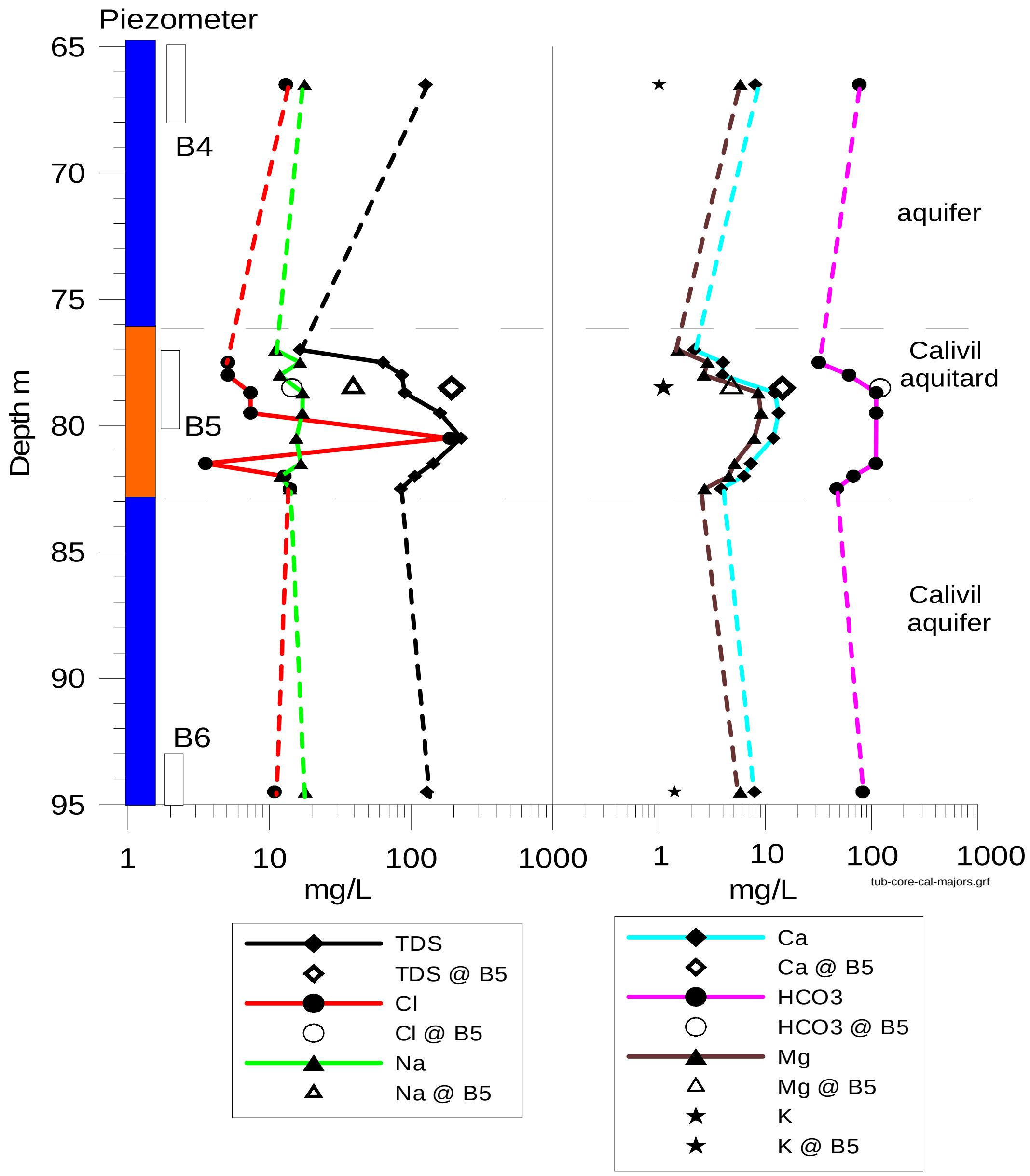


Figure 8.3: Relationship betweeen major ion chemistry of porewater determined from aquitard core samples and groundwater from piezometers within the Calivil aquitard and adjacent aquifers.

8.2 Groundwater water quality degradation related to stress

An association between increasing groundwater salinity and groundwater level drawdown at the Rice Mill and CIA pump sites is evident in Figure 8.4. A broad, but not perfect, correlation is clearly present.

At the Rice Mill site, the most significant increase in shallow groundwater salinity occurred between 1985 and 1994 at a time when there was full seasonal recovery of hydraulic heads, and a slightly increased dynamic range of about 4 m. A further increase in groundwater salinity between 1994 and 1999 correlated with a total drawdown of about 2 m indicating loss of storage over this period. Although groundwater salinity data is scanty, salinity trends appear to have developed a new dynamic equilibrium during the late 1990s during which time dynamic groundwater level range was greatly increased to 12 m seasonally.

Whilst this is strong evidence that groundwater levels are a factor in groundwater salinity increase, close examination of this data is does not support an exclusive cause and effect relationship (Iman & Conover 1983). Other factors which were not monitored, such as the depth to saturated zone, and increased leaching of overlying saline clay may also play a role.

The fact that groundwater salinity increase appears to have occurred prior to major drawdown indicates the extreme sensitivity of groundwater quality. Such a finding implies that numerical models which use only waterlevels to assess the sustainability of groundwater abstraction may not be adequate to assess salinisation risk. Groundwater levels can not be used as a proxy for groundwater quality changes.

Downgraded beneficial use

Increasing salinity at several of the selected sites has resulted in a serious downgrading of beneficial use (Figure 8.5). Prior to 1987, groundwater at all of the selected sites was of potable quality, and suitable for irrigating even the most sensitive crops such as clover. However, by 1995, shallow groundwater at Rice Mill and CIA pump site were no longer potable, and were limited to irrigating more salt tolerant crops such as phalaris and olives. Groundwater salinity at the Darlington Point south site was also increasing significantly.

It is uncertain whether a small decrease in salinity at the two most saline sites in recent years is significant without continued monitoring. The increased frequency of sampling during this period may have captured a natural short term variability, or it is possible that quasi-equilibrium has been achieved between the quantity of salt available

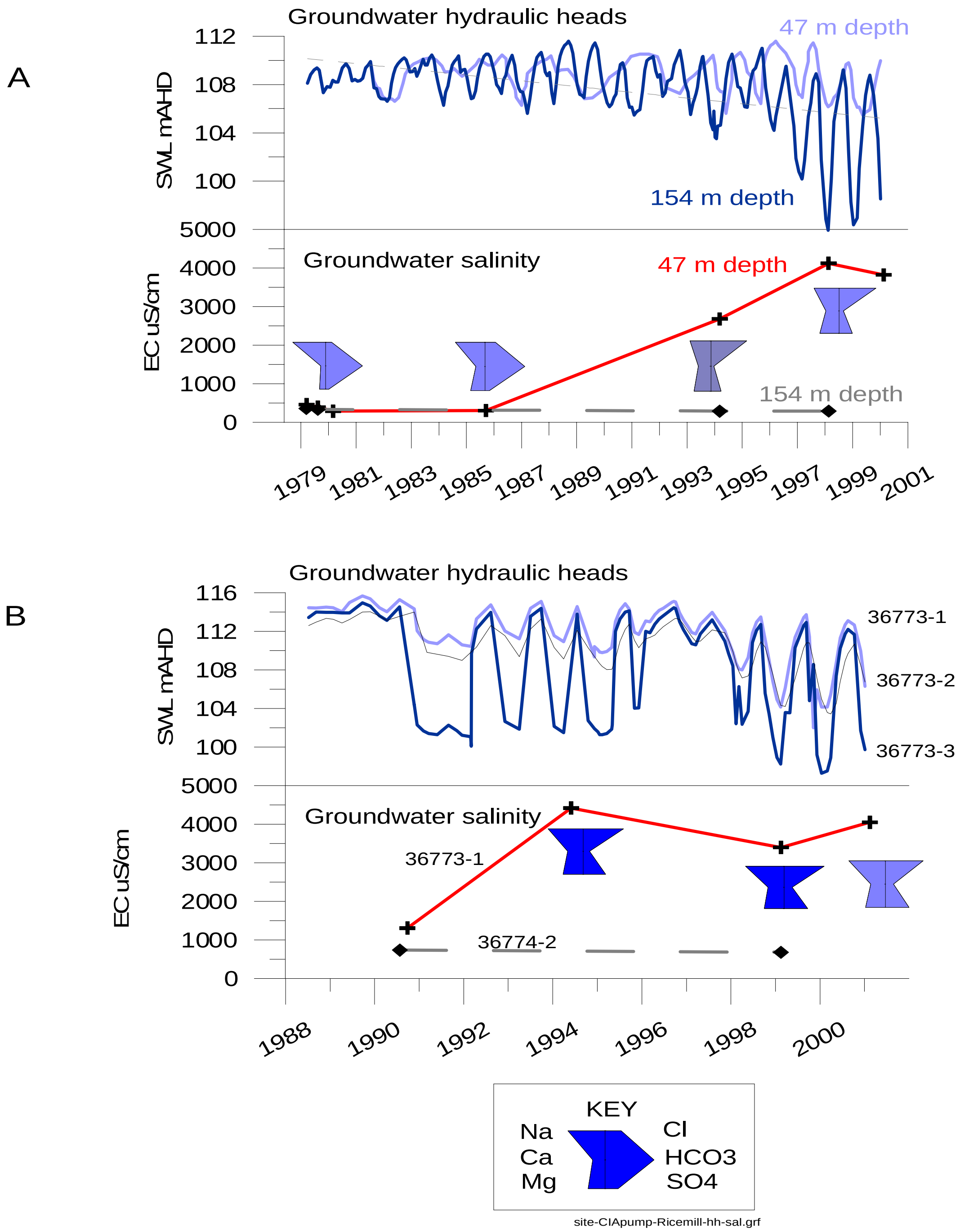


Figure 8.4: Correlation of groundwater level and salinity trends. **A** Rice Mill site, 1979–2001 **B** CIA pump site, 1990–2001.

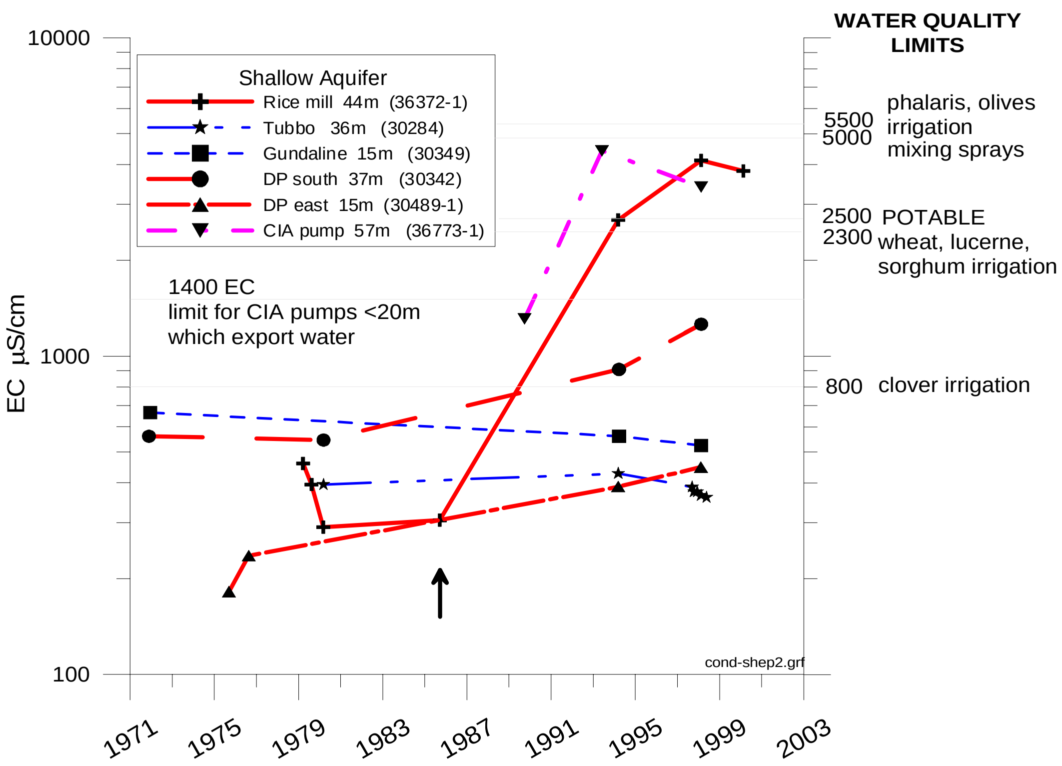


Figure 8.5: Shallow groundwater salinity trends compared with water salinity criteria, 1970–2001

to leach in the alluvial profile, and the capacity of the aquifer to disperse it.

Additional details of groundwater quality impacts with regard to SAR, which is important for resource management, are included on the CD-ROM.

8.3 Hydrochemical verification of flow modelling

Comparison of hydrodynamic and hydrochemical information at the Tubbo site was possible only in a semi-quantitative sense. As reviewed in Chapter 3.6.2, this was to be expected given the complexities of 3D modelling in a stressed system where there are many unknowns (Bath & Lalieux 1997). Lack of direct correlation was partly due to incomplete overlap of hydrodynamic data and hydrochemical data. For instance, hydrochemical changes in aquifer 1 were well defined, though the flow model was not adequately constrained. Also, due to interference effects on groundwater levels, the radial flow model was limited to a 2 week period, although hydrochemical changes were observed over several months.

There was however, semi-qualitative agreement between hydrochemical evidence and leakage estimates derived from the flow model at Tubbo. The upper boundary condition of no recharge during the 15 day period was justified by isotope data. No significant influx of evaporated water was evident in δ^{18}O and δ^{2}H data (Figure 7.10). Also, δ^{13}C was constant between Oct-98 and Feb-99, suggesting that either no groundwater flux occurred during this period, or the more likely scenerio, that influx was characterised by the same isotopic signature (Table 7.7). Groundwater in this aquifer is not modern, as indicated by very low tritium.

Direct mixing between the Lower Shepparton and Calivil aquifers was not able to be tested by hydrogeochemical modelling due to a large advective component that violates the conditions of the mass balance approach. However, both hydrodynamic and hydrochemical data suggest direct mixing of aquifer 3 & 4 could occur through aquitard windows, with little or no aquitard interaction.

Despite these drawbacks, the conjunctive use of hydrodynamic and hydrochemical techniques was of great value in improving the confidence level of estimates for vertical leakage at Tubbo site. Evidence over the long term, and to a lesser degree over the short term demonstrated that vertical leakage did in fact have a measureable impact on groundwater quality.

8.4 Summary

- Groundwater quality of the shallow and middle aquifer has been degraded at some locations beyond beneficial use for irrigation and potable supply purposes. Development of groundwater resources by pumping of the confined aquifer was shown to be a major cause of these effects, on the basis of correlation, consistency, responsiveness and the identification of plausible mechanisms.

- A broad association was evident between groundwater salinity increase and groundwater level drawdown since the mid 1980s and early 1990s at several sites, due to pumping. The lack of direct correlation, however, suggested that factors other than groundwater pumping may also be important.

- Solute transport mechanisms that accounted for variable water quality over time and with depth were identified. Over the short term, advective solute transport appeared to dominate flux through the aquitard between the shallow and middle aquifers, and also through aquitard windows between the middle and deep aquifers. Diffusive flux of solutes to and from the aquitard matrix and fracture flow were of secondary importance.

Chapter 9

CONCLUSION

9.1 Key findings

A summary of key findings was provided at the conclusion of each chapter dealing with hydrogeological, hydrodynamic, and hydrochemical aspects of the research, including the synthesised interpretation provided in the preceeding chapter. However, the pertinent points are revisited here, in terms of the local and international significance of the research findings.

This comprehensive study, employing a wide range of independent analytical tools provides a sound scientific basis for concluding that groundwater development can have a major impact on groundwater quality where vertical leakage occurs rapidly through structures such as aquitard windows. Groundwater quality was observed to be highly sensitive, and in some cases independent of, groundwater level changes, irrespective of the volume of groundwater which was abstracted.

There were a number of key findings that were of significance to the local Lower Murrumbidgee area:

- Aquitard windows may occur, buried beneath a thin veneer of cracking clay, in areas distant from known paleochannels. These aquitard windows occur at a scale which is unlikely to be detected by conventional investigation methods. Lithological logs derived from rotatry drilling operations generally fail to recognise indurated aquitard layers and the density of boreholes is far below that required to adequately define the architecture of mixed-load fluvial deposits (2–4 per km^2).

- Direct evidence for aquitard windows within the Upper Shepparton formation was provided by electrical imaging (120–240 m cross-sections to ∼40 m depth) of bulk electrical conductivity, verified with borehole logs and core analysis. Aquitard windows at depth were detected by delayed hydraulic head changes in specific aquifer-aquitard sequences with the onset of pumping, and hydrochemical changes that were observed due to mixing of groundwater.

- Hydraulic connection between the Upper Shepparton and Lower Shepparton aquifers was variable, depending on the permeability and spatial extent of the indurated and iron-rich clayey sand which forms an effective hydraulic barrier. However, the Lower Shepparton and Calivil aquifer are directly connected by aquitard windows that permit direct groundwater mixing during periods of stress caused by groundwater pumping.

- Isolated nitrate contamination and increased salinity within the Upper and Lower Shepparton aquifers have occured at many sites since the mid-1980s such that TDS and sodium concentrations now limit the beneficial use of some groundwater. However, no corresponding trend was detected over the medium term in the deep Calivil and some minor freshening occured at sites close to the river. It is clear that the groundwater system has responded to development predominately by enhanced vertical leakage.

- During the 1998-99 irrigation season, hydrochemical changes were observed throughout the groundwater system, including the deep Calivil aquifer, after an irrigation bore was activated. A large, but delayed TDS increase was observed in the Upper Shepparton aquifer. Soon after pumping began, a small pulse of saline groundwater in the Lower Shepparton aquifer, and a very small (but significant relative to detection limits) pulse of saline groundwater in the Calivil aquifer were observed. The latter dissipated over a period of several weeks due to lateral influx of fresher water.

- The Calivil aquifer appears to have increased in salinity over a period prior to the first groundwater quality records in the 1970s. Further analysis of pore water diffusion profiles within the Calivil aquitard, may indicate whether this increase occurred with early groundwater development, or is a trend over geologic time.

In summary, key findings that were of significance, beyond the Lower Murrumbidgee in an international context:

- Aquitard windows allow significant downwards advective transport of contaminates when the groundwater system is stressed by development. These landscape features, at a scale of 10s to 100s of metres are not detected by conventional laboratory scale testing or regional scale numerical models. In the absence of a sufficient density of lithological bore logs, innovative field techniques such as electrical imaging and intensive water quality monitoring are required to detect the presence of aquitard windows.

- Vertical fluxes through aquitard windows are likely to dominate leakage via other rapid flow pathways such as fractures and aquifer-aquitard interaction by diffusion. Consequently, in complex stratigraphic environments such as alluvial fans, investigations must be targeted at a scale suitable for detecting aquitard windows.

- The fact that hydraulic heads were not sensitive to increased groundwater flow velocity through aquitard windows confirms recent international studies. Therefore, the standard practice of calibrating flow models to observed hydraulic heads does not quarantee that vertical fluxes of significance for solute transport are accounted for.

- Groundwater chemistry throughout such a complex alluvial aquifer-aquitard system may vary over short periods of time. 'Snapshot' groundwater quality sampling of a large number of bores scattered over a wide area on a limited number of occassions is a necessary first step to identifying contamination issues, but is inadequate to characterise groundwater processes. This research has demonstrated the value of intensive monitoring to establish a cause-effect relationship and the relative importance of factors which contribute to changing groundwater quality.

9.2 Verification of hypotheses and revised conceptual model

Clearly, this research verified both of the hypotheses which were proposed (Section 1.2), and significantly expanded understanding of fundamental groundwater processes in complex alluvial environments. Based on the diverse dataset which was collected and analysed, a series of conceptual models of vertical leakage in an alluvial aquifer-aquitard system under stress have been developed (Figure 9.1). This is a considerable advance over the working conceptual model that was intially proposed (Figure 1.1).

Aquitards in a complex aquifer-aquitard system may be either laterally continuous (system A), fractured (system B) or be punctuated by aquitard windows (system C). Under natural hydraulic gradients, there is little or no downwards flow and diffusion is likely to dominate vertical flux in each system (scenario 1). However, if stress is applied to the system by pumping from the confined aquifer, vertical flux is greater if rapid flow

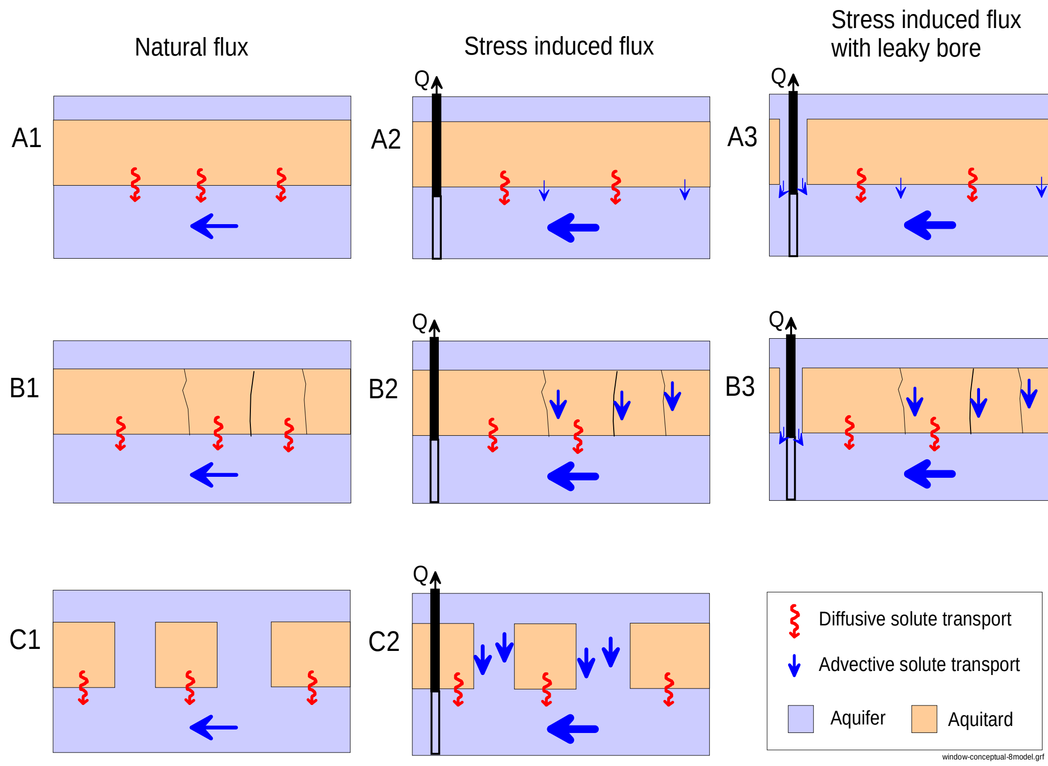


Figure 9.1: Conceptual models of an aquifer-aquitard system depicting the relative importance of vertical leakage and diffusive versus advective solute transport under natural and developed conditions (Flux is proportional to the number and size of arrows). **System A** Continuous aquitard layer. **System B** Fractured aquitard. **System C** Discontinuous aquitard with windows. **Scenario 1** Natural flux. **Scenario 2** Stress induced flux **Scenario 3** Leaky borehole flux.

pathways exist (scenario 2). Advective flux is greatest through aquitard windows, although leaky boreholes (scenario 3) may also be important.

The relative importance of fracturing and leaky boreholes will depend on the density and interconnectiveness of these features. However even narrow rapid flow pathways may significantly alter the groundwater velocity field. In the absence of sufficient borehole lithological data, a realistic conceptual model of an aquifer-aquitard system may be identified by monitoring depth-specific groundwater pressure changes and hydrochemical variability during a pump test.

9.3 Lessons learnt from this research

As anticipated, this research project has generated a tool kit of techniques and principles that are readily transferrable to similar environments in the Murray-Darling Basin and beyond. So that the experience gained during this project is of benefit to other workers, a few comments are offered regarding what worked and what didn't.

- An integrated approach utilising geological, geophysical, geotechnical, hydrochemical, isotopic and numerical modelling techniques was of immense value in characterising realistic groundwater processes, and greatly increased confidence in results.
- Simultaneous monitoring of water level and water quality during a pump test generated a valuable dataset. However, more frequent water quality monitoring (eg. daily for a week), augmented by continuous downhole measurement of key parameters, is advisable if logistics and finances permit.
- An air-conditioned and well equiped field laboratory greatly facililated the acquisition of timely, accurate and precise hydrochemical data.
- The additional expense of advanced isotope analysis techniques, which eliminated complex field proceedures requiring large volumes of water, was fully justified by enhanced accuracy and reduced fieldwork costs.
- Hydrogeochemical modelling suffered from a lack of regional context and mixing endmembers. For example, there is virtually no data available for the Renmark aquifer, and spatially variability within specific aquitards.
- The opportunity to gain a tremendous amount of information about the groundwater system (eg. relative hardness and permeability of each layer) was lost by not being on site during drilling, coring and piezometer installation.
- More reliable and time efficient estimates of core permeability could have beeen obtained using a triaxial cell with back pressure, or centrifuge modelling, rather than a simplified falling head device.

9.4 Future challenges

While this work clearly has implications for groundwater management, specific actions that are required to sustain the beneficial use of groundwater in accordance with the NSW State Groundwater Quality Protection Policy (DLWC 1998b) are beyond the scope of this thesis and will be suggested in a separate report. Challenges for future investigations and research will now be outlined.

Vertical leakage pathways in the upper aquitard and indurated clayey sand.

- The relative importance of various leakage pathways through the semi-saturated upper aquitard remains largely unknown, while there is some uncertainty regarding important solute transport mechanisms through the indurated clayey sand.

- Large aquitard windows were detected in an area that had been designated as an interchannel deposit, indicating that further work is required to define the spatial distribution these deposits and entrained salt. The nonlinear relationship between salt content and bulk EC measured by geophysical techniques requires further evaluation.

- The overall effect of significant rapid leakage pathways which have been introduced through the upper aquitard depends on an unknown number of bores which have corroded or were gravel packed to the surface during construction.

- The indurated clayey sand aquitard has been identified as a relatively effective hydraulic barrier between the shallow and deep aquifers, yet its spatial extent and dependance of permeability on groundwater chemistry have not been determined.

Hydrogeochemical interactions with clayey aquitards and aquifer matrix.

- Improved understanding of variable groundwater chemistry requires not only a more representative regional dataset that includes adjacent aquifers and spatial variability within aquitards, but information on important hydrogeochemical processes in the system.

- Sodium concentrations over the short and long term are evidently controlled by ion exchange with clay, although the cation exchange capacity of the various clay deposits to buffer salinity changes in the system have not been determined.

- Groundwater residence times have been brought into question by inexplicable isotope results which suggest complex clay matrix interaction. The reason why aquitard pore water appeared to be younger than groundwater in an adjacent aquifer remains a mystery.

- Predictive groundwater models are yet to be developed which incorporate both rapid flow pathways and reactive contaminate transport that accounts for matrix interaction.

9.5 Conclusion

Although a large quantity of groundwater is stored within the alluvial deposits of the Lower Murrumbidgee, there is evidence that groundwater quality is highly sensitive to agricultural development. Short and long term groundwater quality changes were associated with downwards leakage through rapid flow pathways. The pathways are dominantly aquitard windows and leaky bores and are instigated as the result of stressing the hydraulic system by abstraction from the underlying confined aquifer.

Where leakage pathways exist in the vadose zone, groundwater quality change may also be caused by drainage of irrigation water. Such changes in water and contaminant flux may occur prior to, or independently of decreased hydraulic pressure caused by stressing an underlying aquifer.

As is the nature of science, future investigation will, it is hoped, further improve (and perhaps modify elements of this thesis) our understanding of groundwater within the earth underfoot.

Personal communications

Dr. R.I. Acworth, Groundwater Centre, UNSW Water Research Laboratory, Manly Vale, NSW, 2093.

J.R.W. Bell, Groundwater Centre, UNSW Water Research Laboratory, Manly Vale, NSW, 2093.

J. Gaston, Tubbo Irrigation, Narrandera, NSW, 2700.

M. Groskops, UNSW Water Research Laboratory, Manly Vale, NSW, 2093.

P. Kumar, NSW Department of Land and Water Conservation, Leeton, NSW, 2705.

S. Lawson, NSW Department of Land and Water Conservation, Leeton, NSW, 2705.

J. Lot, Caringal Station, Narrandera, NSW, 2700.

Dr. N. Merrick, National Centre for Groundwater Management, UTS, Ultimo, NSW, 2007.

G. Piscopo, NSW Department of Land and Water Conservation, Parramatta, NSW, 2124.

C. Taylor, formerly at Institute for Geological and Nuclear Sciences, Lower Hutt, New Zealand.

L. Webb, formerly at NSW Department of Land and Water Conservation, Leeton, NSW, 2705.

Dr. B. Williams, 171 Briglow St., Lynham, ACT, 2601.

D. Yu, UNSW Geography, Kensington, NSW, 2052.

Bibliography

Acworth, R. & Beasley, R. (1998). Investigation of EM31 Anomolies at Yarramanbah/Pump Station Creek on the Liverpool Plains of New South Wales. WRL Research Report No. 195.

Acworth, R. I. (1999). Investigation of dryland salinity using the electrical image method. *Australian Journal of Soil Research 37*, 623–636.

Acworth, R. I. & Jankowski, J. (1997). The relationship between bulk electrical conductivity and dryland salinity in the Narrabri Formation at Breeza, Liverpool Plains, New South Wales, Australia. *Hydrogeology Journal 5(3)*, 109–123.

Alekin, O. (1970). *Principles of Hydrochemistry.* Hydromet. Izdat.

Allison, G., Holmes, J., & Hughes, M. (1971). Tritium fallout in southern Australia and its hydrological implications. *Journal of Hydrology 14*, 307–321.

Anderson, M. P. & Woessner, W. W. (1992). *Applied groundwater modeling - Simulation of Flow and Advective Transport.* Academic Press.

Appelo, C. A. J. (1996). Multicomponent ion exchange and chromatography in natural systems. In P. Lichtner, C. I. Steefel, & E. H. Oelkers (Eds.), *Reactive transport in porous media* (Reviews in mineralogy ed.), Volume 34.

Appelo, C. A. J. & Postma, D. (1996). *Geochemistry, groundwater and pollution.* Rotterdam: Balkema.

Arad, A. & Evans, R. (1987). The hydrogeology, hydrochemistry and environmental isotopes of the Campaspe River aquifer system, north-central Victoria, Australia. *Journal of Hydrology 95*, 63–86.

ARMCANZ (1995). Groundwater quality protection. Agriculture and Resource Management Council of Australia and New Zealand.

AS (1990). Test pumping of water wells. Australian Standard AS-2368. pp. 40.

Ayers, J. & Gosselin, D. (1995). Design of a multi-level monitoring well for continuous sample collection. *Groundwater Monitoring Review Fall*, 119–124.

Back, W. (1985). Geochemical significance of aquitards. In *Hydrogeology of Rocks of Low Permeability. Proceedings*, Volume Volume XVII, Tuscon, Arizona. IAH Memoires. 39.

Back, W. (1986). Role of aquitards in hydrogeochemical systems: a synopsis. *Applied Geochemistry 1*, 427–437.

Back, W. & Baedecker, M. (1989). Chemical hydrogeology in natural and contaminated environments. *Journal of Hydrology 106*, 1–28.

Back, W., Baedecker, M., & Wood, W. (1993). Scales in chemical hydrogeology: a historical perspective. In W. M. Alley (Ed.), *Regional ground-water quality.*, pp. 111–1129. New York: Van Nostrand Reinhold.

Ball, J. & Nordstrom, D. (1991). WATEQ4F - User's manual with revised thermodynamic data based and test cases for calculating speciation of major, trace and redox elements in natural waters. U.S. Geological Survey Open-File Report 90-129.

Bardsley, E. & Campbell, D. (1999). Confined aquifers can serve a giant weighing lysimeters. In *Water99 Joint Congress*, Brisbane, 6-8 July. 328.

Bardsley, W. E. & Campbell, D. I. (1994). A new method for measuring near-surface moisture budgets in hydrological systems. *Journal of Hydrology 154*, 245–254.

Barrash, W. & Dougherty, M. E. (1997). Modeling axially symmetric and nonsymmetric flow to a well with MODFLOW, and application to Goddard2 well test, Boise, Idaho. *Groundwater 35*(4), 602–611.

Bath, A. & Lalieux, P. (1997). Use of hydrochemical information in testing groundwater flow models. In NEA-OECD (Ed.), *Technical Summary and Proceedings of a Workshop organised by NEA Co-ordinating Group on Site Evaluation and Design of Experiments for Radioactive Waste Disposal.*, Borgholm, Sweden.

Bauld, J., Sandstrom, M., & Stadter, F. (1995). Australian groundwater quality assessment project: Background and current status. In *Murray-Darling Basin Groundwater Workshop Extended Abstracts*, Wagga Wagga, 11-13 September.

Beattie, J. (1970). Peculiar features of soil development in na deposits in the eastern Riverina, NSW. *Australian Journal of Soil Research 8*, 145–156.

Beck, R. (1999). Procedures for measurements for field and laboratory use. UNSW Water Research Laboratory.

Beke, G. J., Entz, T., & Graham, D. P. (1993). Long-term quality of shallow ground water at irrigated sites. *Irrigation Science 119*(1), 116–128.

Belitz, K. & Bredehoeft, J. (1990). Role of confining layers in controlling large-scale regional ground-water flow. In S. P. Neuman & I. Neretnieks (Eds.), *Hydrogeology of Low Permeability Environments* (International Association of Hydrogeologists Selected Papers ed.), Volume 2. Verlag Heinz Heise.

Belitz, K. & Phillips, S. P. (1995). Alternative to agricultural drains in California's San Joaquin Valley: Results of a regional-scale hydrogeologic approach. *Water Resources Research 31(8)*, 1845–1862.

Biot, M. (1941). General theory of three-dimensional consolidation. *Journal of Applied Physics 12*, 155–164.

Bishop, P. K. (1990). Precipitation of dissolved carbonate species from natural waters for C-13 analysis - a critical appraisal. *Chemical Geology 80*, 251–259.

Bjerg, P. L. & Christensen, T. H. (1992). Spatial and temporal small-scale variation in groundwater quality of a shallow aquifer. *Journal of Hydrology 131*, 133–149.

Bouwer, H. (1987). Effect of irrigated agriculture on groundwater. *Journal of Irrigation and Drainage Engineering 113*(1), 2–15.

Bowles, J. E. (1978). *Engineering properties of soils and their measurement.* New York: McGraw-Hill.

Bowmer, K. H., Korth, W., Scott, A., McCorkelle, G., & Thomas, M. (1998). Pesticide monitoring in the irrigation areas of south-western NSW, 1990-1995, CSIRO technical report 17/98.

Boyce, J. & Eyles, N. (2000). Architectural element analysis applied to glacial deposits: internal geometry of a Late Pleistocene till sheet, southern Ontario, Canada. *Geological Society of America Bulletin* (112), 98–118.

Bredehoeft, J. D., Neuzil, C., & Milly, P. (1983). Regional flow in the dakota aquifer: a study of the role of confining layers. U.S. Geological Survey Water-Supply Paper 2237.

Bredehoeft, J. D. & Papadopulos, S. S. (1980). A method for determining the hydraulic properties of tight formations. *Water Resources Research 16(1)*, 233–238.

Brobst, R. B. & Muszka, P. M. (1986). The effect of three drilling fluids on groundwater sample chemistry. *Groundwater Monitoring Review Winter 1986*, 62–70.

Brown, C. M. (1989). Structural and stratigraphic framework of groundwater occurrence and surface discharge in the Murray Basin, southeastern Australia. *BMR Journal of Australian Geology and Geophysics 11*, 127–146.

Brown, C. M. & Radke, B. M. (1989). Stratigraphy and sedimentology of mid-Tertiary permeability barriers in the subsurface of the Murray Basin, southeastern Australia. *BMR Journal of Australian Geology and Geophysics 11*, 367–385.

Brown, K., Love, A., & Harrington, G. A. (2001). Vertical groundwater recharge to the Tertiary confined sand aquifer, south east South Australia. South Australian Detment for Water Resources. DWR2001/002. pp. 42.

Brownbill, R. (1997). Pumping tests of the calivil formation/renmark group aquifer in the victorian riverine plain: Estimation of vertical leakage and influence on watertable levels. In *Murray-Darling Basin Groundwater Workshop,*, Toowoomba.

Butler, B. E. (1950). A theory of prior streams as a causal factor of soil occurrence in the Riverine Plain of south-eastern Australia. *Australian Journal of Agricultural Research 1*, 231–252.

Butler, B. E. (1976). Subplasticity in Australian Soils: Introduction. *AJSR 14*, 225–6.

Butler, B. E. & Hutton, J. (1956). na in the Riverine Plain of south-eastern Australia and the soils thereon. *Australian Journal of Agricultural Research 7*, 536–553.

Butler, J. J. (1998). *The design, performance and analysis of slug tests.* Lewis.

Calf, G. E. (1987). Tritium activity in Australian Rainwater 1962-1986. Australian Nuclear Science and Technology Organisation. pp. 11.

Cantor, L., Knox, R., & Fairchild, D. (1987). *Ground water quality protection.* Lewis Publishers.

Chamley, H. (1989). *Clay Sedimentology.* London: Springer-Verlag.

Chapelle, F. H. (1993). Geochemical modeling as a tool for studying microbial processes in ground-water systems. In *Ground-water microbiology and geochemistry*, pp. 264 – 294. John Wiley and Sons.

Charman, P. & Murphy, B. (Eds.) (1991). *Soils, their properties and management. A soil conservation handbook for New South Wales.* Sydney University Press.

Chen, X., Ayers, J. F., & Gosselin, D. C. (1998). Analysis of nitrate-nitrogen movement near high-capacity irrigation wells. *Groundwater Monitoring Review Fall 1998*, 148–156.

Christen, E., Prasad, A., & Khan, S. (2000). Annual net recharge analysis for CIA using piezometric data. CSIRO Land and Water and Coleambally Irrigation Limited.

Christen, E., Prasad, A., & Khan, S. (2001). Spatial analysis of shallow groundwater pumping for salinity control and potential conjunctive use: A case study of the Coleambally Irrigation Area. CSIRO Land and Water Technical Report 40-00.

CIC (2000). Environment report. Coleambally Irrigation Corporation Ltd. pp. 101.

Clark, I. & Fritz, P. (1997). *Environmental Isotopes in Hydrogeology.* Lewis Publishers.

Close, M. (1989). Effect of serial correlation on ground water quality sampling frequency. *Water Resources Bulletin 25*(3), 507–515.

Close, M. E. (1987). Effects of irrigation on water quality of a shallow unconfined aquifer. *Water Resources Bulletin 23(5)*, 793–803.

Collis-George, N. & Yates, D. B. (1987). The effects of encapsulated air on falling head permeaters. *Soil Science 143(5)*, 318–328.

Cosler, D. (1997). Groundwater sampling and time-series evaluation techniques to determine vertical concentration distributions. *Groundwater 35*(5), 825–841.

Craig, H. (1957). Isotopic standards for carbon and oxygen and correction factors for mass-spectrometric analysis of carbon dioxide. *Geochemica et Cosmochimica Acta 12*, 133–149.

Craig, H. (1961). Isotopic variations in meteoric waters. *Science 133*, 1702–1703.

Crawford, R. L. (1997). *Salinity on the Liverpool Plains, NSW, Australia: Hydraulic conductivity and mobility of salt in smectitie clays.* Fourth year thesis, unpublished, University of New South Wales.

Custodio, E. (1991). Hydrogeological and hydrochemical aspects of aquifer overexploitation. In I. Simmers, F. Villarroya, & L. F. Rebollo (Eds.), *Selected Papers on Aquifer Overexploitation from the 23rd International Congress of IAH, Puerto de la Cruz, Tenerife (Spain), April 15-19, 1991*, Volume 3, pp. 3–26. International Association of Hydrogeologists.

Dansgaard, W. (1964). Stable isotopes in precipitation. *Tellus 16*, 436–468.

D'Astous, A. Y., Ruland, W. W., Bruce, J. R. G., Cherry, J. A., & Gillham, R. W. (1989). Fracture effects in the shallow groundwater zone in weathered Sarnia-area clay. *Canadian Geotechnical Journal 26*, 43–56.

Desaulniers, D. E., Cherry, J. A., & Fritz, P. (1981). Origin, age and movement of pore water ion argillaceous quaternary deposits at four sites in southwestern Ontario. *Journal of Hydrology 50*, 231–257.

Desaulniers, D. E., Cherry, J. A., & Fritz, P. (1982). Origin, age, and movement of pore water in clayey pleistocene deposits in south central Canada. In: Isotopes Studies of Hydrologic Processes. Ed: Perry, EC, Montgomery, CW, (1982). Selected papers from a symposium held Sept. 19,1980 at Northern Illinois University.

Desbarats, A., Hinton, M., & Logan, C. (2001). Geostatistical mapping of leakance in a regional aquitard, Oak Ridges Moraine area, Ontario, Canada. *Hydrogeology Journal 9*, 79–96.

Diersch, H. (1998). FEFLOW user's manual - interactive, graphics-based finite element simulation system for modeling groundwater flow, contaminate mass and heat transport processes. WASY Institute for Water Resources Planning and Systems Research Ltd.

DLWC (1998a). Aquifer risk assessment report. NSW Detment of Land and Water Conservation.

DLWC (1998b). The NSW State groundwater quality protection policy. NSW Detment of Land and Water Conservation.

Domenico, P. & Schwartz, F. (1990). *Physical and chemical hydrogeology.* New York: John Wiley and Sons.

Doser, L., Longstaffe, F., Ferrell, R., & Walthall, P. (1998). Fluid flow through clayey soils: stable isotope and mineralogical evidence. *Clay Minerals 33*, 43–49.

Drever, J. I. & Smith, C. L. (1978). Cyclic wetting and drying of the soil zone as an influence on the chemistry of groundwater in arid terrains. *American Journal of Science 278*, 1448–1454.

Drury, L. W. (1982). Age, movement and recharge of groundwater in Tertiary sediments of the Lower Murrumbidgee River Valley. Water Resources Commission NSW, Hydrogeological Report No 1982-11.

Drury, L. W., Calf, G. E., & Dharmasiri, J. K. (1984). Radiocarbon dating of groundwater in Tertiary sediments of the eastern Murray Basin. *Australian Journal of Soil Research 22*, 379–387.

DWR (1993). Water resources of the Murrumbidgee Valley. NSW Detment of Water Resources.

Edmunds, W. M. (1994). Characterization of groundwaters in semi-arid and arid zones using minor elements. In H. Nash & G. J. M. McCall (Eds.), *Groundwater Quality* (17th Special Report ed.)., pp. 19–30. London: Chapman and Hall.

Enever, D. (1999). Coleambally groundwater simulation model. CSIRO Land and Water Consultancy Report to Coleambally Irrigation.

Evans, R. S. (1989). Saline water disposal options in the Murray Basin. *BMR Journal of Australian Geology and Geophysics 11*, 167–185.

Evans, W. & Bauld, J. (1993). Towards an Australian groundwater quality assessment program. *AGSO Journal of Australian Geology and Geophysics 14*(2/3), 307–311.

Evans, W. R. & Kellett, J. R. (1989). The hydrogeology of the Murray Basin, southeastern Australia. *BMR Journal of Australian Geology and Geophysics 11*, 147–166.

Fetter, C. W. (1993). *Contaminant Hydrogeology*. New York: Maxwell Macmillan Internatonal.

Fetter, C. W. (1994). *Applied Hydrogeology* (3rd ed.). New Jerzy: Prentice Hall.

Firman, J. (1994). Paleosols in laterite and silcrete profiles - evidence from the south east margin of the Australian Precambrian Shield. *Earth Science Reviews 36*, 149–179.

Flossman, R. (1999). The preation of thin polished and double polished thin sections. [online], Available : http://www.geology.unsw.edu.au/staff/flossman/sectioning.html [2000, September 13].

Fogg, G. (1986). Groundwater flow and sand body interconnectedness in a thick multiple-aquifer system. *Water Resources Research 22*(5), 679–694.

Freeze, R. & Cherry, J. (1979). *Groundwater*. Prentice Hall.

Fried, A. (1993). Late Pleistocene river morphological change, southeastern Australia: the conundrum of sinuous channels during the Last Glacial Maximum. *Paleogeography, Palaeoclimatology, Palaeoecology 101*, 305–316.

Galloway, W. E. & Hobday, D. K. (1996). *Terrigenous Clastic Depositional Systems: Applications to Fossil Fuel and Groundwater Resources.* (2nd ed.). Springer.

Garrels, R. & Mackenzie, F. (1967). Origin of the chemical composition of springs and lakes. In *Equilibrium Concepts in Natural Water Systems, Advances in Chemisty Series*, Volume 67, pp. 222–242. Washington, D.C.: American Chemistry Society. cited by khurst 97.

Gerber, R., Boyce, J., & Howard, K. (2001). Evaluation of heterogeneity and field-scale groundwater flow regime in a leaky till aquitard. *Hydrogeology Journal 9*, 60–78.

Gerber, R. & Howard, K. (2000). Recharge through a regional till aquitard: three-dimensional flow model water balance approach. *Groundwater 38*(3), 410–422.

Ghassemi, F., Alam, K., & Howard, K. (2000). Fresh-water lenses and practical limitations of their three-dimensional simulation. *Hydrogeology Journal 8*, 521–537.

Ghassemi, F., Jakeman, A., & Nix, H. (1995). *Salinisation of Land and Water Resources: Human causes, extent, management and case studies.* Sydney: University of New South Wales Press.

Gorelick, S. (1983). A review of distributed ameter groundwater management modeling methods. *Water Resources Research 19*(2), 305–319.

Gribble, C. & Hall, A. (1985). *A practical introduction to optical mineralogy.* Chapman and Hall.

Grisak, G. & Cherry, J. (1975). Hydrologic characteristics and response of a fractured till and clay confining a shallow aquifer. *Canadian Geotechnical Journal 12*, 23–43.

HACH (1989). *Water Analysis Handbook.* Lovedale, Colorado: HACH Company.

Hallows, P. J. (1987). The extent and effects of shallow water tables in NSW irrigated areas. In *Australian Water Resources Council Conference Series, Proceedings of the International Conference on Groundwater Systems Under Stress*, Brisbane, 1986, pp. 31–42.

Hanor, J. S. (1987). Report on rates and mechanisms of vertical fluid flow and solute transport across the Geera aquitard, central Murray Basin, southeastern Australia. Unpublished report, Bureau of Mineral Resources.

Hantush, M. (1960). Modification of the theory of leaky aquifers. *Journal of Geophysical Research 65*, 3713–3725.

Hantush, M. & Jacob, C. (1955). Non-steady radial flow in an infinite leaky aquifer. *Transactions American Geophyisical Union 36*, 95–100.

Hare, P. W. (1997). Water-level fluctuations due to barometric pressure changes in an isolated portion of an unconfined aquifer. *Groundwater 35*(4), 667–671.

Harrington, G., Love, A., & Herczeg, A. (2001). Relative importance of physical and chemical processes affecting solute distributions in a clay aquitard. In R. Cidu (Ed.), *Proceedings of the 10th International Symposium on Water-Rock Interaction,*, Villasimius, Italy, pp. 177–180. Balkema.

Harris, J., Loftis, J., & Montgomery, R. (1987). Statistical methods for characterizing ground-water quality. *Groundwater 25*(2), 185–193.

Harrison, B., Sudicky, E. A., & Cherry, J. A. (1992). Numerical analysis of solute migration through fractured clayey deposits into underlying aquifers. *Water Resources Research 28(2)*, 515–526.

Helsel, D. & Hirsch, R. (1992). *Statistical methods in water resources.* Studies in Environmental Science 49. Amsterdam: Elsevier.

Hem, J. (1992). Study and interpretation of the chemical characteristics of natural water. USGS Water Supply Paper 2254.

Hendry, M. J. (1982). Hydraulic conductivity of a glacial till in alberta. *Groundwater 20*(2), 162–169.

Hendry, M. J. & Schwartz, F. W. (1988). An alternative view on the origin of chemical and isotopic patterns in groundwater from the Milk River aquifer, Canada. *Water Resources Research 24(10)*, 1747–1763.

Hendry, M. J. & Wassenaar, L. I. (1999). Implications of the distribution of delta-D in pore waters for groundwater flow and the timing of geologic events in a thick aquitard system. *Water Resources Research 35*(6), 1751–1760.

Hendry, M. J. & Wassenaar, L. I. (2000). Controls on the distribution of major ions in pore waters of a thick surficial aquitard. *Water Resources Research 36(2)*, 503–513.

Hibbs, B. J. (1998). The city of El Paso/Ciudad Juarez Corridor international case study of groundwater depletion and aquifer salinization. In J. V. Brahan (Ed.), *Gambling with Groundwater - Phyiscal, Chemical and Biological Aspects of Aquifer-Stream Relations*, pp. 359–367. St.Paul, Minnesota: American Institute of Hydrology.

Hillel, D. (1991). *Out the Earth: Civilisation and the Life of the Soil.* New York: Maxwell Macmillan.

Holtz, R. D. & Kovacs, W. D. (1981). *An introduction to geotechnical engineering.* Prentice-Hall.

Hornbuckle, J. & Christen, E. (1999). Physical properties of soils in the Murrumbidgee and Coleamablly Irrigation Areas. CSIRO Land and Water, Griffith, Technical Report 17/99.

Hounslow, A. (1995). *Water Quality Data: Analysis and Interpretation.* New York: CRC Lewis.

Hoyle, B. (1989). Ground-water quality variations in a silty alluvial soil aquifer, Oklahoma. *Groundwater 27(4)*, 540–549.

Hubble, G. D. (1981). The cracking clay soils: definition, distribution, nature, genesis and use. In J. W. McGarity, E. H. Hoult, & H. B. So (Eds.), *The properties and utilization of cracking clay soils.* Reviews in Rural Science, Proceedings of a symposium at the University of New England, NSW, 24-28 August 1981.

Iman, R. & Conover, W. (1983). *A modern approach to statistics.* New York: John Wiley and Sons.

Jacobson, G., Jankowski, J., & Abell, R. S. (1991). Groundwater and surface water interaction at Lake George, New South Wales. *BMR Journal of Australian Geology and Geophysics 12*, 161–190.

Jankowksi, J. & Beck, P. (2000). Aquifer heterogeneity: hydrogeological and hydrochemical properties of the Botany Sands aquifer and their impact on contaminant transport. *Australian Journal of Earth Sciences 47*, 45–64.

Jankowski, J., Acworth, I., & Little, W. (1994). The hydrogeology and hydrogeochemical processes associated with the occurrence of dryland salinity in the Strontian Road Basin, NSW. In *Water Down Under '94*, Water Down Under '94, Adelaide, Australia, 21-25 November, 1994, Adelaide, 21-25 November 1994.

Jiwan, J. & Gates, G. (1995). Assessment of pesticides in groundwater in the Upper Namoi. Detment of Water Resources, Technical Services Division.

Johnson, A. (1967). Specific yield - compilation of specific yields for various materials. U.S. Geological Survey Water-Supply Paper 1662-D.

Johnson, R. L., Cherry, J. A., & Pankow, J. F. (1989). Diffusive contaminant transport in natural clay: a field example and implications for clay-lined waste disposal sites. *Environmental Science and Technology 23*, 340–349.

Jones, B. F., Hanor, J. S., & Evans, W. R. (1994). Sources of dissolved salts in the central Murray Basin, Australia. *Chemical Geology 111*, 135–154.

Keary, J. (1993). National groundwater protection guidelines for Australia. *AGSO Journal of Australian Geology and Geophysics 14*(2/3), 303–306.

Keith, S., Wilson, L., Fitch, H., & Esposito, D. (1983). Sources of spatial-temporal variability in ground-water quality data and methods of control. *Groundwater Monitoring Review 1983*(Spring), 21–32.

Keller, C. K., van der Kamp, G., & Cherry, J. A. (1986). Fracture permeability and groundwater flow in clayey till near Saskatoon, Saskatchewan. *Canadian Geotechnical Journal 23*, 229–240.

Keller, C. K., Van der Kamp, G., & Cherry, J. A. (1988). Hydrogeology of two Saskatchewan tills I. Fractures, bulk permeability, and spatial variability of downward flow. *Journal of Hydrology 101*, 97–121.

Keller, C. K., Van der Kamp, G., & Cherry, J. A. (1989). A multiscale study of the permeability of a thick clayey till. *Water Resources Research 25(11)*, 2299–2317.

Keller, C. K., van der Kamp, G., & Cherry, J. A. (1991). Hydrogeochemistry of a clayey till 1. spatial variability. *Water Resources Research 27(10)*, 2543–2554.

Kelly, I. (1978). Geochemistry of ground waters in the Riverine Plain. In R. Storrier & I. Kelly (Eds.), *The hydrogeology of the Riverine Plain of South East-Australia*, Griffith, NSW, July 28-29, 1978. Australian Society of Soil Science, Riverina Branch.

Khan, S., Merrick, N. P., Christen, E., & Tiwari, A. (1999). Review of regional groundwater model of Coleambally Irrigation Area. In *Murray-Darling Basin Groundwater Workshop, September 1999*, Griffith.

Khan, S., Short, L., & Best, L. (2000). Groundwater dynamics in the Coleambally Irrigation Area. CSIRO Land and Water Consultancy Report to Coleambally Irrigation Corporation.

Kloppmann, W., Matray, J., & Aranyossy, J. (2001). Contamination of deep formation waters by drilling fluids: correction of the chemical and isotopic composition and evaluation of errors. *Applied Geochemistry 16*, 1083–1096.

Kolditz, O., Ratke, R., Hans-Jorg, G., & Zielke, W. (1998). Coupled groundwater flow and transport: 1. verfication of variable density flow and transport models. *Advance in Water Resources 21*(1), 27–46.

Konikow, L. F. & Arevalo, J. (1993). Advection and diffusion in a variable-salinity confining layer. *Water Resources Research 29*(8), 2747–2761.

Konikow, L. F. & Person, M. (1985). Assessment of long-term salinity changes in an irrigated stream aquifer system. *Water Resources Research 21(11)*, 1611–1624.

Kruseman, G. P. & de Ridder, N. A. (1990). *Analysis and evaluation of pumping test data* (2nd ed.). Netherlands: ILRI Publication 47.

Lacombe, S., Sudicky, E. A., Frape, S. K., & Unger, A. J. A. (1995). Influence of leaky boreholes on cross-formational groundwater flow and contaminant transport. *Water Resources Research 31(8)*, 1871–1882.

Lambe, T. W. (1951). *Soil Testing for Engineers.* New York: John Wiley & Sons Ltd.

Landmeyer, J. E. (1996). Aquifer response to record low barometric pressures in the Southeastern United States. *Groundwater 34(5)*, 917–924.

Lawrence, C. (1975). Geology, hydrodynamics and hydrochemistry of the sourthern Murray Basin. Geological Survey of Victoria Memoir 30.

Lawson, S. (1992). *Groundwater behaviour in the confined aquifers of the Murrumbidgee alluvial fan.* M.Sc. thesis, unpublished, University of New South Wales.

Lawson, S. (1996). *Review of groundwater use and water level behaviour in the Lower Murrumbidgee Valley: Groundwater Status Report No. 4.* Detment of Land and Water Conservation, Murrumbidgee Region, Technical Report No. 96/01.

Lawson, S. & Holland, G. F. (1999). Leakage between shallow and deep aquifers in the southeastern Murray Basin. In *Murray Darling Basin Groundwater Workshop, Integrated Perspectives, 14-16th September, 1999*, Griffith.

Lawson, S. & van der Lely, A. (1992). *Coleambally Deep Bore Project: Analysis of long term pumping test.* Detment of Water Resources, Technical Services, Murrumbidgee Region.

Lawson, S. & Webb, E. (1998). Review of groundwater use and groundwater level behaviour in the Lower Murrumbidgee Valley. Detment of Land and Water Conservation, Murrumbidgee Region, Technical Report No. 98/05.

Lewis, D. W. & McConchie, D. (1994). *Analytical sedimentology.* New York: Chapman and Hall.

Loftis, J. C. (1996). Trends in groundwater quality. *Hydrological Processes 10*, 335–355.

Love, A. J., Herczeg, A. L., & Walker, G. (1995). Transport of water and solutes across a regional aquitard inferred from porewater deuterium and chloride profiles. Vienna, Austria, pp. 273–286. Isotopes in Water Resources Management, IAEA.

Macumber, P. (1991). *Interaction between groundwater and surface systems in Northern Victoria.* East Melbourne, Australia: Detment of Conservation and Environment.

Maloszewski, P. & Zuber, A. (1991). Influence of matrix diffusion and exchange reactions on radiocarbon ages in fissured carbonate aquifers. *Water Resources Research 27*(8), 1937–1945.

Martin, P. & Frind, E. (1998). Modeling a complex multi-aquifer system: the waterloo moraine. *Ground Water 36*(4), 679–690.

Matthes, G., Munnich, K. O., & Sonntag, C. (1976). Practical problems of groundwater model ages for groundwater protection studies. In *Interpretation of environmental isotope and hydrochemical data in groundwater hydrology. Proceedings of an advisory group meeting organized by the International Atomic Energy Agency, Vienna, 27-31 January, 1975*, Vienna. International Atomic Energy Agency.

Maurer, D. K., Johnson, A. K., & Welch, A. H. (2000). Hydrogeology and potential effects of changes in water use, Carson Desert Agricultural Area, Churchill County, Nevada. US Geological Survey Water-Supply Paper 2436.

McCloskey, T. F. & Finnemore, E. J. (1996). Estimating hydraulic conductivities in an alluvial basin from sediment facies models. *Groundwater 34*(6), 1024–1032.

McDonald, M. & Harbaugh, A. (1988). MODFLOW: A modular three-dimensional finite-difference groundwater flow model. USGS Scientific Publications.

McKay, L. D., Cherry, J. A., & Gillham, R. W. (1993). Field experiments in a fracture clay till 1. Hydraulic conductivity and fracture aperture. *Water Resources Research 29(4)*, 1149–1162.

MCM (2001). Murrumbidgee catchment management, Murrumbidgee Catchment Management Board, [online], Available: http://www.murrumbidgee-catchment.org.au [2001, Nov 20].

McNally, G. & Sutherland, M. (1997). Engineering geology of prior stream deposits on the Riverine Plain, eastern Murray Basin, NSW. In G. McNally (Ed.), *Collected Case Studies in Engineering Geology, Hydrogeology and Environmental Geology*, pp. 361. Environmental, Engineering and Hydrogeology Specialist Group EEHSG of the Geological Society of Australia.

McNeill, J. (1986). Geonics EM-39 borehole conductivity meter: theory of operation. Geonics Ltd., Technical Note TN-20.

Meyer, W. S. (1992). *Sustainability of land and water resources used for Australian Irrigated Agriculture: a research strategy postition paper.* CSIRO Division of Water Resources, Series No. 8.

MGMC. Management of the alluvial groundwater resources of the Lower Murrumbidgee Valley, NSW: A discussion paper. Murrumbidgee Groundwater Management Committee.

Montgomery, R., Loftis, J., & Harris, J. (1987). Statistical characteristics of ground-water quality variables. *Groundwater 25*(2), 176–184.

Moody, D. W., Carr, W., Chase, E. B., & Paulson, R. W. (1986). National water summary 1986 - hydrologic events and groundwater quality. US Geological Survey Water-Supply Paper 2325.

MSU (2000). Development of flow and transport models, Moscow State University, [online], Available: http://www.geol.msu.ru/deps/hydro/grants/crdf/index.htm [2000, May 19].

Mundorff, M., Carrigan, P., Steele, T., & Randall, A. (1976). Hydrologic evaluation of salinity control and reclamation projects in the indus basin, pakistan - a summary. U.S. Geological Survey Water-Supply Paper 1608-Q.

Nelson, J. & Ward, R. (1981). Statistical considerations and sampling techniques for ground-water quality monitoring. *Groundwater 19*(6), 617–625.

Neuman, S., Fogg, G., & Jacobson, E. (1980). A statistical approach to the inverse problem of aquifer hydrology. *Water Resources Research 16*(1), 32–58.

Neuman, S., Preller, C., & Narasimhan, T. (1982). Adaptive explicit-implicit quasi three-dimensional finite element model of flow and subsidence in multiaquifer systems. *Water Resources Research 18*(5), 1551–1561.

Neuman, S. & Witherspoon, P. (1969a). Applicability of current theories of flow in leaky aquifers. *Water Resources Research 5*(4), 817–829.

Neuman, S. & Witherspoon, P. (1969b). Theory of flow in a confined two aquifer system. *Water Resources Research 5*(4), 803–816.

Neuman, S. & Witherspoon, P. (1972). Field determination of the hydraulic properties of leaky multiple aquifer systems. *Water Resources Research 8*, 1284–1298.

Neuman, S. P. & Gardner, D. A. (1989). Determination of aquitard/aquiclude hydraulic properties from arbitary water-level fluctuations by deconvolution. *Groundwater 27*(1), 66–77.

Neuzil, C. E. (1982). On conducting the modified "slug" test in tight formations. *Water Resources Research 18(2)*, 439–441.

Neuzil, C. E. (1986). Groundwater flow in low-permeability environments. *Water Resources Research 22(8)*, 1163–1195.

Neuzil, C. E. (1994). How permeable are clays and shales ? *Water Resources Research 30(2)*, 145–150.

NFF (1996). Australian agriculture. National Farmers Federation.

Nightingale, H. & Bianchi, W. (1980). Well-water quality changes correlated with well pumping time and aquifer ameters - Fresno, California. *Groundwater 18*(3), 274–280.

Nightingale, H. I. & Bianchi, W. C. (1977). Ground-water chemical quality management by artificial recharge. *Groundwater 15(1)*, 15–22.

Northcote, K. H., Hubble, G. D., Isbell, R. F., Thompson, C. H., & Bettenay, E. (1975). *A Description of Australian Soils*. CSIRO.

OECD (1998). Sustainable management of water in agriculture: Issues and policies. Organisation for Economic Development.

Ortega-Guerrero, A., Cherry, J. A., & Aravena, R. (1997). Origin of pore water and salinity in the lucustrine aquitard overlying the regional aquifer of Mexico City. *Journal of Hydrology 197*, 47–69.

Page, K. J., Nanson, G. C., & Price, D. M. (1991). Thermoluminescence chronology of Late Quaternary deposition on the Riverine Plains of south-eastern Australia. *Australian Geographer 22*(1), 14–23.

Page, K. L., Nanson, G., & Price, D. (1996). Chronology of Murrumbidgee River palaeochannels on the Riverine Plain, southeastern Australia. *Quaternary Science 11(4)*, 311–326.

Page, K. L. & Nanson, G. C. (1996). Stratigraphic architecture resulting from Late Quaternary evolution of the Riverine Plain, south-eastern Australia. *Sedimentology 43*, 927–945.

Pain, C. & Ollier, C. (1996). Regolith stratigraphy: principles and problems. *AGSO Journal of Australian Geology and Geophysics 16*(3), 197–202.

Pearson, F. (1965). Use of C-13/C-12 ratios to correct radocarbon ages of material initially diluted by limestone. In *Proceedings of the 6th International Conference on Radiocarbon and Tritium Dating*, Pulman, Washington, pp. 357. cited by Clark and Fritz 97.

Pels, S. (1964). Quaternary sedimentation by prior streams on the Riverine Plain, south-west of Griffith, NSW. *Journal and Proceedings of the Royal Society of NSW. 97*, 107–115.

Pels, S. (1968). Environmental studies of the Coleambally Irrigation Area, and surrounding districts t I, Geology and ground water hydrology. Water Conservation and Irrigation Commission Land Use Series 2:3.

Pettyjohn, W. (1976). Monitoring cyclic fluctuations in groundwater-water quality. *Groundwater 14*(6), 472–479.

Pettyjohn, W. (1982). Cause and effect of cyclic changes in ground-water quality. *Groundwater Monitoring Review* (Winter), 43–49.

Piscopo, G. (1997). Groundwater vulnerability Murrumbidgee Catchment: map and notes. Detment of Land and Water Conservation, Hydrogeology Unit.

Plummer, L., Jones, B., & Truesdell, A. (1976). WATEQF - A FORTRAN IV version of WATEQ, a computer program for calculating chemical equilibria of natural waters. USGS Water Resources Investigations Report 76-12.

Plummer, L., Rupert, M., Busenberg, E., & Schlosser, P. (2000). Age of irrigation water in groundwater from the Eastern Snake River Plain Aquifer, South-Central Idaho. *Groundwater 38*(2), 264–283.

Plummer, L. N. (1984). Geochemical modeling: a comison of forward and inverse methods. In B. Hitchon & E. Wallick (Eds.), *Proceedings 1st Canadian American Conference on Hydrogeology, Practical Applications of Ground Water Chemistry*, Worthington, Ohio, pp. 149–177. National Groundwater Well Association. 301.

Prasad, A., Christen, E., & Khan, S. (2001). The potential role for deep groundwater pumping in the control of irrigation induced salinity in the Riverine Plain: A case study of the Coleambally Irrigation Area. CSIRO Land and Water Technical Report 01-01.

Pratha, S. A., Lawson, S., & Enever, D. J. (1997). *Hydrogeology of the Coleambally Irrigation Area: a brief description for use with a groundwater simulation model.* CSIRO Land and Water Technical Report 3/97.

Pucci, A. A., Ehlke, T., & Owens, J. (1992). Confining unit effects on water quality in the New Jersey coastal plain. *Groundwater 30*(3), 415–42.

Punthakey, J., Pratha, S., Somaratne, N., Merrick, N., Lawson, S., & Williams, R. (1995). Modelling impacts of regional development and environmental change in the Lower Murrumbidgee River Basin. In P. Binning, H. Bridgman, & B. Williams (Eds.), *MODSIM International Congress on Modelling and Simulation*, Volume 3, University of Newcastle, pp. 29–34.

Punthakey, J., Somaratne, N., Pratha, S., Merrick, N., Lawson, S., & Williams, R. (1994). Regional groundwater modelling of the Lower Murrumbidgee River Basin. Detment of Water Resources.

Pye, K. (1983). Red beds. In A. Goudie & K. Pye (Eds.), *Chemical sediments and geomorphology: precipitates and residua in the near-surface environment.*, pp. 439. Academic Press.

Rajagopal, R. (1986). The effect of sampling frequency on ground-water quality characterization. *Groundwater Monitoring Review 1986*(Fall), 65–73.

Ramamurthy, L. M. & Holmes, J. W. (1983). Detection of subsurface seepage between aquifers by hydrochemical and environmental isotopic techniques - a case study from South Australia. In *Proceedings of the Hamburg Symposium Relation of Groundwater Quality and Quantity, August, 1983*, Hamburg. IAH Publication No. 146. 184.

Rao, N. (1998). Impact of clayey soils on nitrate pollution in the groundwater of the lower Vamsadhara River basin, India. *Hydrological Sciences 43*(5), 701–714.

Rasmussen, T. C. & Crawford, L. A. (1997). Identifying and removing barometric pressure effects in confined and unconfined aquifers. *Groundwater 35(3)*, 502–511.

Reilly, T. E. & Harbaugh, A. (1993). Simulation of cylindrical flow to a well using the U.S. Geological Survey Modular Finite-Difference Ground-Water Flow Model. *Ground Water 31*(3), 489–494.

Reimann & Caret, D. (1998). *Chemical elements in the environment.* Springer Verlag.

Remenda, V. (2001). Theme issue on confining units. *Hydrogeology Journal 9*, 3–4.

Remenda, V. H. & van der Kamp, G. (1997). Contamination from sand-bentonite seal in monitoring wells installed in aquitards. *Groundwater 35*(1), 39–46.

Rojstaczer, S. (1988). Determination of fluid flow properties from the response of water levels in wells to atmospheric loading. *Water Resources Research 24(110)*, 1927–1938.

Ronald, R. (1960). *The Riverina, Peoples and Properties.* Cheshire, Melbourne.

Ronen, D., Magaritz, M., Gvirtzman, H., & Garner, W. (1987). Microscale chemical heterogeneity in groundwater. *Journal of Hydrology 92*, 173–178.

Rowe, R. K., Quicgley, R. M., & Booker, J. R. (1995). *Clayey barrier systems for waste disposal facilities.* London: E & FN Spon.

Rozanski, K., Araguas-Araguas, L., & Gonfiantini, R. (1993). Isotopic patterns in modern global precipitation. In *Continental Isotope Indicators of Climate.* American Geophysical Union Monograph. cited by Clark and Fritz 1997.

Rudolph, D. L., Cherry, J. A., & Farvolden, R. N. (1991). Groundwater flow and solute transport in fractured lacustrine clay near Mexico City. *Water Resources Research 27(9)*, 2187–2201.

Rudolph, D. L. & Frind, E. O. (1991). Hydraulic response of highly compressible aquitards during consolidation. *Water Resources Research 27(1)*, 17–30.

Sanford, W. (1997). Correcting for diffusion in carbon-14 dating of groundwater. *Groundwater 35*(2), 357–361.

Schmidt, K. (1977). Water quality variations for pumping wells. *Groundwater 15*(2), 130–137.

Schmidt, K. D. & Sherman, I. (1987). Effect of irrigation on groundwater quality in California. *Irrigation Science 113(1)*, 16–29.

Schmidt, L., Telfer, A., & Waters, M. (1996). Pesticides and nitrate in groundwater in relation to landuse in the south east of South Australia. Detment of Environment and Natural Resources.

Schumm, S. (1968). River adjustment to altered hydrologic regimen - Murrumbidgee River and palaeochannels, Australia. Geological Society of America Professional Paper.

Sepulveda, N. & Zack, A. L. (1991). The effects of overburden stress on the specific storage and hydraulic conductivity of artesian aquifers. *Journal of Hydrology 128*, 305–321.

Short, L. & Khan, S. (2000). Quantifying climatic impacts on shallow watertables in the CIA. CSIRO Land and Water and Coleambally Irrigation Limited.

Simpson, H. J. & Herzceg, A. L. (1991). Stable isotopes as an indicator of evaporation in the River Murray, Australia. *Water Resources Research 27(8)*, 1925–1935.

Singh, B. & Narain, P. (1980). Seasonal fluctuations in the quality of underground irrigation water in a brackish water affected tract. *Agrochemica Vol XXIV No. 2-3*, 169–175.

Skempton, A. W. (1970). The consolidation of clays by gravitational compaction. *Quarterly Journal Geological Society London 125*, 373–411.

SKM (1995). *Girgarre Salinity Control Project. Goulburn-Murray Water.* Tatura, Victoria: Sinclair Knight Merz.

SKM (1997). *Pumping Test of the Calivil Formation/Renmark Group Aquifer: Campaspe West and Murray Valley. Goulburn-Murray Water.* Tatura, Victoria: Sinclair Knight Merz.

Smith, D. I. (1999). *Water in Australia: Resources and Management.* Melbourne: Oxford University Press.

Smith, L. & Schwartz, F. (1981). Mass transport, 3, Role of hydraulic conductivity data in prediction. *Water Resources Research 17*(5), 1463–1479.

Snow, D. (1969). Anisotropic permeability of fractured media. *Water Resources Research 5*(6), 1273–1289.

Sophocleous, M. & Ma, T. (1998). A decision support model to assess vulnerability to salt water intrusion in the Great Bend Prairie aquifer of Kansas. *Groundwater 36*(3), 476–483.

Spalding, R. & Exner, M. (1980). Areal, vertical and temporal difference in ground water chemistry. I Inorganic constituents. *Journal of Environmental Quality 9*(3), 466–479.

Spears, D. A. & Reeves, M. J. (1975). The influence of superficial deposits on groundwater quality in the Vale of York. *Quarterly Journal of Engineering Geology 8*, 255–269.

Stadter, F. & Love, A. J. (1989). The Tatiara proclaimed region, southeastern South Australia: hydrogeological investigations and groundwater management. *BMR Journal of Australian Geology and Geophysics 11*, 253–259.

Stannard, M. (1978). The effect of irrigation on groundwaters in the mid-Murrumbidgee region. In *The hydrogeology of the Riverine Plain of South East-Australia.*

Stephans, D. B. (1983). Groundwater flow and implications for groundwater contamination north of Prewitt, New Mexico, USA. *Journal of Hydrology 61*, 391–408.

Stigter, T. Y., van Ooijen, S. P. J., Post, V. E. A., Appelo, C. A. J., & Carvalho Dill, A. M. M. (1998). A hydrogeological and hydrochemical explanation of the groundwater composition under irrigated land in a Mediterranean environment, Algarve, Portugal. *Journal of Hydrology 208*, 262–279.

Stow, D. A. (1981). Fine-grained sediments: Terminology. *Quarterly Journal of Engineering Geology 14*, 243–244.

Strack, O. D. L. (1981a). Flow in aquifers with clay laminae 1. The comprehensive potential. *Water Resources Research 17(4)*, 985–992.

Strack, O. D. L. (1981b). Flow in aquifers with clay laminae 2. Exact solutions. *Water Resources Research 17(4)*, 993–1004.

Strack, O. D. L. & Haitjema, H. (1981a). Modelling double aquifer flow using a comprehensive potential and distributed singularities. 1 Solution for homogeneous permeability. *Water Resources Research 17*(5), 1535–1549.

Strack, O. D. L. & Haitjema, H. (1981b). Modelling double aquifer flow using a comprehensive potential and distributed singularities. 2 Solution for inhomogeneous permeabilities. *Water Resources Research 17*(5), 1551–1560.

Streltsova, T. (1976). Hydrodynamics of groudwater flow in a fractured formation. *Water Resources Research 12*(3), 405–414.

Stuiver, M. & Polach, H. (1977). Reporting of 14C data. *Radiocarbon 19*, 355–363.

Sudicky, E. & Frind, E. O. (1981). Carbon 14 dating of groundwater in confined aquifers: implications of aquitard diffusion. *Water Resources Research 17*(4), 1060–1064.

Sukhija, B. S., Reddy, D. V., & Nagabhushanam, P. (1998). Environmental tracers and hydrodynamics of over-exploited aquifers. In e. a. Brahana (Ed.), *Gambling with Groundwater - Physical, Chemical and Biological Aspects of Aquifer-Stream Relations*, pp. 521–526.

Swan, A. & Sandilands, M. (1995). *Introduction to geological data analysis*. Blackwell Scientific.

Talsma, T. & Flint, S. (1958). Some factors determining the hydraulic conductivity of subsoils with special reference to tile drainage problems. *Soil Science 85*, 198–206.

Talsma, T. & van der Lelij, A. (1976). Infiltration and water movement in an in situ swelling soil during prolonged ponding. *Australian Journal of Soil Research 14*, 337–349.

Tamers, M. (1975). The validity of radiocarbon dates on groundwater. *Geophysical Survey 2*, 217–239.

Timms, W. (1997). *Liverpool Plains Water Quality Project 1996/97: Report on Groundwater Quality*. Detment of Land and Water Conservation, CNR97.108.

Timms, W. (1998). Hydraulic linkages between shallow saline groundwaters and pressurised alluvial aquifers. Detment of Land and Water Conservation, Centre for Natural Resources, Research Report T2816.

Timms, W., Acworth, I., & Berhane, D. (2001). Shallow groundwater dynamics in smectite dominated clay on the Liverpool Plains of New South Wales. *Australian Journal of Soil Research 39*(2), 203–218.

Timms, W., Acworth, I., Jankowski, J., & Lawson, S. (1999). Groundwater quality in a aquifer aquitard system subjected to large volume abstraction for irrigation in the Lower Murrumbidgee. In *Murray-Darling Basin Groundwater Workshop "Integrated Perspectives"*, Griffith, pp. 131–136.

Timms, W. & Acworth, R. (2002a). Induced leakage due to groundwater pumping and flood irrigation: Pullaming Agricultural Trial Site, Liverpool Plains. UNSW Water Research Laboratory, Research Report No. 208. Available: http://www.wrl.unsw.edu.au/ research.

Timms, W. & Acworth, R. (2002b). Origin, lithology and weathering characteristics of Late Tertiary Quaternary aquitards on the Lower Murrumbidgee alluvial fan. *Australian Journal of Earth Sciences. 49*, 525–537.

Timms, W., Acworth, R., Jankowksi, J., & Lawson, S. (2000). Groundwater quality trends related to aquitard salt storage at selected sites in the Lower Murrumbidgee alluvium, australia. In O. Sililo (Ed.), *Proceedings of the XXX International Association of Hydrogeologists Congress "Groundwater: Past Achievements and Future Challenges". 26 November - 1 December 2000*, Cape Town, South Africa, pp. 655–660.

Timms, W., Acworth, R., & Jankowski, J. (2001). Quantifying leakage and mixing in an alluvial aquifer system: a combined hydrochemical and hydrodynamic modelling approach constrained by isotopic evidence. In K. Seiler & S. Wohnlich (Eds.), *Proceedings of the XXXI International Association of Hydrogeologists Congress "New Approaches Characterizing Groundwater Flow", 10-14 September, 2001*, Munich, Germany, pp. 419–423. Balkema.

Timms, W., Acworth, R., Young, R., & Berhane, D. (2002). Natural leakage pathways through cracking smectite clay: Hudson Agricultural Trial Site, Liverpool Plains. UNSW Water Research Laboratory, Research Report No. 209. Available: http://www.wrl.unsw.edu.au/ research.

Trewhella, W. (1989). Irrigation recharge. *BMR Journal of Australian Geology and Geophysics 11*, 187–194.

USGS (2001). Saline intrusion along preferrential flowpaths in a multi-layered aquifer, Atlantic coast, US. U.S. Geological Survey [online], Available: http://water.usgs.gov/ogw/gwrp/saltwater/fig5.html [2001, September 13].

van der Kamp, G. (2001). Methods for determining the in situ hydraulic conductivity of shallow aquitards - an overview. *Hydrogeology Journal 9*, 5–16.

van der Kamp, G. & Gale, J. (1983). Theory of earth tide and barometric effects in porous formations with compressible grains. *Water Resources Research 19*(2), 538–544.

van der Kamp, G. & Keller, C. K. (1993). Casing leakage in monitoring wells: Detection, confirmation, and prevention. *Groundwater Monitoring Review Fall*, 136–141.

van der Kamp, G. & Maathuis, H. (1991). Annual fluctuations of groundwater levels as a result of loading by surface moisture. *Journal of Hydrology 127*, 137–152.

van der Lelij, A. (1989). Groundwater accessions in the Coleambally irrigation area of NSW. In Sharma (Ed.), *Groundwater Recharge, Proceedings of the symposium on groundwater recharge*, Mandurah, 1987. Balkema, Rotterdam.

van der Lelij, A. (1990). Watertable behaviour as a function of rice growing and the resulting salinity hazard. Detment of Water Resources, Revised 1980 Technical Report.

van der Lelij, A., Ellis, L., & Gee, B. (1987). Groundwater hydrology of the Coleambally Irrigation Area. Detment of Water Resources, Technical Report.

van der Lely, A. (2000). Salinity survey results, Coleambally Irrigation Area 1999. Detment of Land and Water Conservation.

van Dijk, D. C. & Talsma, T. (1964). Soils of portion of the Coleambally Irrigation Area, NSW. CSIRO Soils and Land Use Series.

Velde, B. (1995). *Origin and Mineralogy of Clay: Clays and the Environment.* Springer.

Walker, G. R. & Cook, P. G. (1991). The importance of considering diffusion when using carbon-14 to estimate groundwater recharge to an unconfined aquifer. *Journal of Hydrology 128*, 41–48.

Wallbrink, P., Murray, A., & Olley, J. (1998). Determining sources and transit times of suspended sediment in the Murrumbidgee River, New South Wales, Australia, using fallout 137Cs and 210Pb. *Water Resources Research 34*(4), 879–887.

Wassenaar, L. I. & Hendry, M. J. (2000). Mechanisms controlling the distribution and transport of 14C in a clay-rich till aquitard. *Groundwater 38*(3), 343–349.

Wassenaar, L. L. & Hendry, M. J. (1999). Improved piezometer construction and sampling techniques to determine pore water chemistry in aquitards. *Groundwater 37*(4), 564.

Weber, K. & van Geuns, L. (1990). Framework for constructing clastic reservoir simulation models. *Journal of Petroleum Technology 43*, 1248–1253, 1296–1297.

Williams, B., Walker, J., & Tane, H. (2001). Drier landscapes and rising watertables: an ecohydrological adox. *Natural Resource Management 4*(1), 10–18.

Williams, B. G. & Ward, J. K. (1987). The chemistry of shallow groundwaters in the Murrumbidgee Irrigation Area, New South Wales. *Australian Journal of Soil Research 25*, 251–261.

Williams, B. G., Ward, J. K., & Belbin, L. (1987). Numerical classification of saline groundwater chemistry in the Murrumbidgee Irrigation Area. *Australian Journal of Soil Research 25*, 337–345.

Williams, R. M. (1987). The Murray Basin aquifer system with some reference to developing salinity problems. In *International Conference on Groundwater Systems Under Stress*, Brisbane, pp. 311–320. Australian Water Resources Council Conference Series.

Wilmoth, B. (1972). Salty ground water and meteoric flushing of contaminated aquifers in West Viginia. *Groundwater 10*(1), 99–105.

Wilson, M. (1999). The origin and formation of clay minerals in soils: past, present and future perspectives. *Clay Minerals 34*, 7–25.

Wood, W., Rainwater, K., & Thompson, D. (1997). Quantifying macropore recharge: examples from a semi-arid area. *Ground Water 35*(6), 1097–1106.

Woolley, D. (1991). Narranderra hydrogeological map (1:250 000 scale). Bureau of Mineral Resources Geology and Geophysics.

Woolley, D. & Kalf, F. (1981). Groundwater resources of the Lower Lachlan and Murrumbidgee Valleys. In *9th Convention of AWWA, 6-10th April, 1981*, Perth. cited by Williams RM 1987.

Woolley, D. & Williams, R. (1978). Tertiary stratigraphy and hydrogeology of the eastern t of the Riverine Plain. In R. Storrier & I. Kelly (Eds.), *The Hydrogeology of the Riverine Plain, ASSS Conference Proceedings*, Griffith, NSW.

WRC (1984). Groundwater in New South Wales.

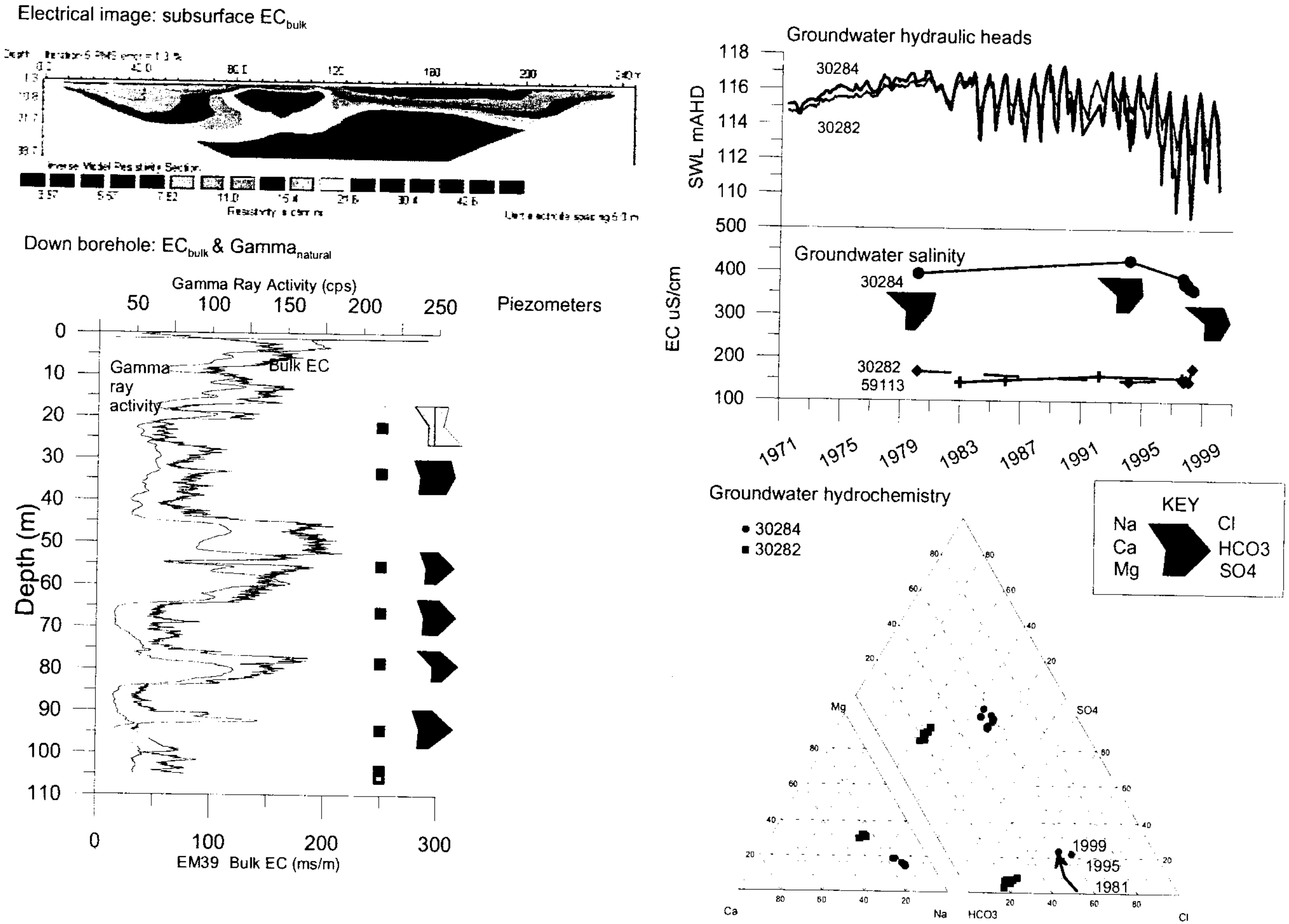
Younger, P. L. (1992). The hydrogeological use of thin sections: inexpensive estimates of groundwater flow and transport ameters. *Quarterly Journal of Engineering Geology 25*, 159–164.

Zheng, C. (1990). MT3D a modular three-dimensional transport model. S.S. Papadopulos & Associates, Inc.

Appendix A

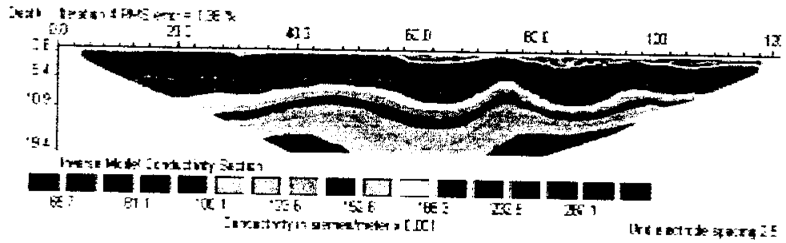
Site summaries of key geophysical and hydrochemical data

TUBBO SITE

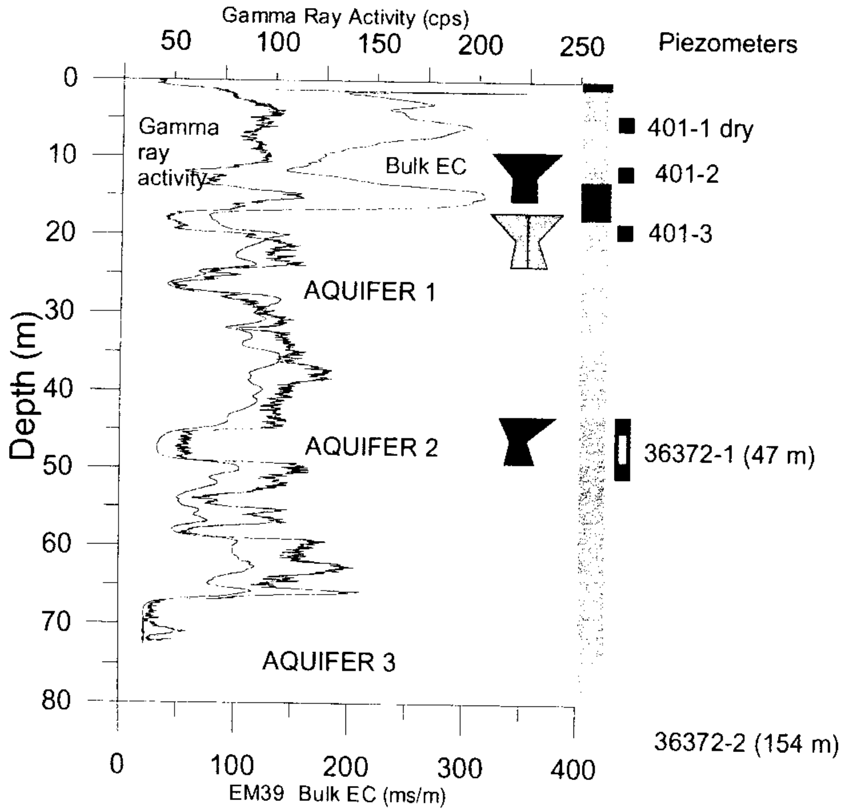

Electrical image: subsurface EC bulk
Inverse Model Resistivity Section
Resistivity in ohm.m
Down borehole: EC bulk & Gamma natural
Gamma Ray Activity (cps)
Piezometers
Gamma ray activity
Bulk EC
Depth (m)
EM39 Bulk EC (ms/m)
Groundwater hydraulic heads
SWL mAHD
30284
30282
Groundwater salinity
EC uS/cm
59113
1971
1975
1979
1983
1987
1991
1995
1999
Groundwater hydrochemistry
30284
30282
KEY
Na
Ca
Mg
Cl
HCO3
SO4
1981

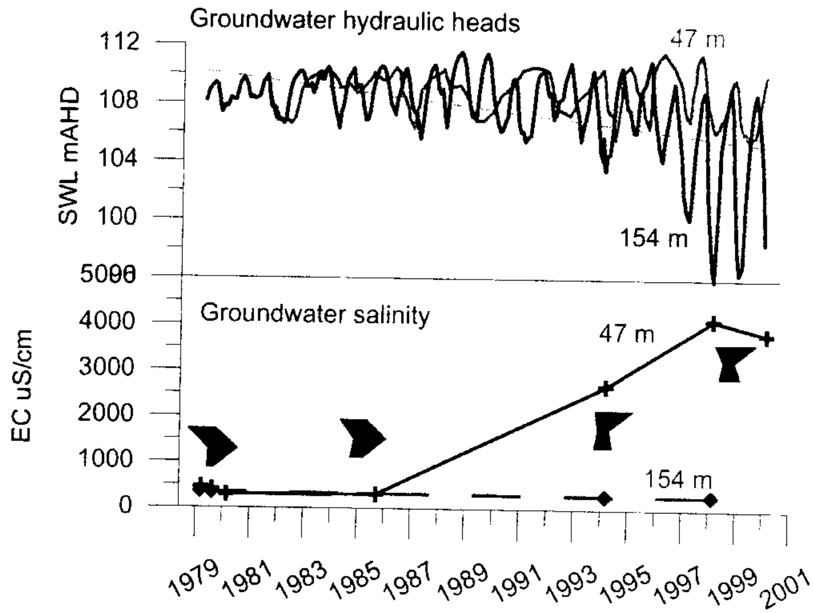
RICE MILL SITE

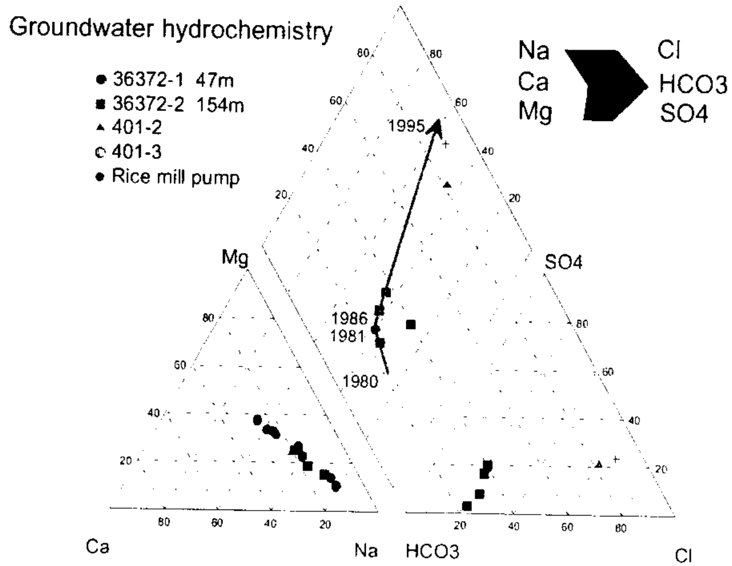
Electrical image: subsurface EC_{bulk}



Down borehole: EC_{bulk} & $Gamma_{natural}$







Appendix B

Bore and piezometer details

Table B.1: Bore installation details for bores sampled.

Bore	Type [1]	Screen depth (m)	Total depth (m)	Easting	Northing	Elev. [2] (m AHD)	Install date	Install method	Casing ID (mm)
			-	442550	6166550	-			
RIVER									
Rice Mill site									
36372-1	M	44-50	200	404480	6154900	123.46	Mar-80	RM	102
36372-2	M	151-157	200	404480	6154900	123.91	Mar-80	RM	102
36372-3	M	188-194	200	404480	6154900	123.83	Mar-80	RM	102
401-1	M	4.5-6	20	404480	6154900	-	Nov-00	RM	50
401-2	M	11-12.5	20	404480	6154900	-	Nov-00	RM	50
401-3	M	18.5-20	20	404480	6154900	-	Nov-00	RM	50
33804	D	30.4-31.5	31.6	404465	6155050	-	Feb-72	-	127
CIA pump site									
39406	I	74-79,81-108,119-134	-	409850	6144050	-	Aug-88	-	590
12986	M	9	9	410480	6143780	-	-	-	50
12985	M	10	10	410480	6143780	-	-	-	50
36774-1	M	49-59	154	410480	6143780	126.33	May-88	RM	100
36773-1	M	57-63	178	409745	6144275	126.8	May-88	RM	100
36773-2	M	122-128	178	409745	6144275	126.76	May-88	RM	100
36774-2	M	136-148	154	409745	6144275	126.32	May-88	RM	100
36773-3	M	166-172	178	409745	6144275	126.78	May-88	RM	100
Darlington Point east									
30489-1	M	15.2-19.4	115.8	416487	6167017	128.96	Aug-74	RM	102
30489-2	M	82.2-84.4	115.8	416487	6167017	129.16	Aug-74	RM	64
30489-3	M	101.2-110.3	115.8	416487	6167017	129.65	Aug-74	RM	102
Darlington Point south									
30342	M	37.4-43.4	-	404500	6167760	124.47	Nov-72	RM	127
30341	M	121-129	134	404500	6167760	124.46	Nov-72	RM	127
Gundaline									
30349	M	15.2-18.2	19.8	383100	6176150	116.86	Dec-72	RM	101
A1	M	47.5	72	383100	6176150	-	Jun-98	RM	50
A2	M	72	72	383100	6176150	-	Jun-98	RM	50
30348	M	128-134	140.2	383100	6176150	116.83	Dec-72	RM	101

(continued on next page)

[2]casing top

[2]I is irrigation, M is monitoring, D is domestic

Table B.1: (continued)

Bore	Type	Screen depth (m)	Total depth (m)	Easting	Northing	Elev. (m AHD)	Install date	Install method	Casing ID (mm)
Irrigation bores									
65947	I	151.1-325.3[3]	407	431832	6161180	-	Aug-78	CT	1200
59202	I	80.8-133.8[4]	135.6	434220	6158745	139	Apr-85	CT	406
400694	I	44-90[5]	90	435495	6160800	-	Jul-98	R	224
400066	I	86-111.4[6]	127	417940	6158540	-	-	R	-
60487	I	71.6-137.5[7]	138.7	401490	6166755	-	Nov-85	CT	-
400029	I	85.9-129.5[8]	131	429875	6163280	-	Jun-96	R	610
400064	I	87.8 - 147.8[9]	150	418010	6152790	-	Mar-96	R	450
Tubbo site									
B1	M	21-24	80	427435	6161840		Jun-98	RM	50
B2	M	32-35	69	427435	6161840		Jun-98	RM	50
30284	M	36.2 -39.5	45.4	429180	6161900	133.66	May-72	RM	203
B3	M	54-57	96	427435	6161840		Jun-98	RM	50
B4	M	65-68	69	427435	6161840		Jun-98	RM	50
B5	M	77-80	80	427435	6161840		Jun-98	RM	50
B6	M	93-96	96	427435	6161840		Jun-98	RM	50
59113	I	86.8-120.7[10]	123.2	427410	6161865	131	Jan-84	CT	406
30282	M	100.2-117.2[11]	117.6	429180	6161900	133.75	May-72	RM	117
CHAN	[12]	-	-			-	-	-	-

[3] (7 intervals)
[4] (8 intervals)
[5] (3 intervals)
[6] (10 intervals)
[7] (5 intervals)
[8] (7 intervals)
[9] (7 intervals)
[10] (5 intervals)
[11] (2 intervals)
[12] at outlet of Turkey's nest near 59113

Appendix C

Enclosed CD-ROM "Publications & Datasets"

C.1 Publications and reporting

Research seminar presented at CSIRO Land & Water, 11th July, 2001 *(.pdf file)*
Poster - "Sustainable pumping below saline clay....." *(.pdf)*
Poster - "Aquifer salinisation, Lower Murrumbidgee Australia" *(.pdf)*

Timms, W., Acworth, R.I. (2002b). Origin, lithology and weathering characteristics of Late Tertiary-Quaternary clay aquitard units on the Lower Murrumbidgee alluvial fan. *Australian Journal of Earth Sciences* 49:525-537 *(.pdf)*

Timms, W., Acworth, R.I. and Jankowski, J., 2001. Quantifying leakage and mixing in an alluvial aquifer system: a combined hydrochemical and hydrodynamic modelling approach constrained by isotopic evidence. *In: K.P. Seiler and S. Wohnlich (Editors), Proceedings of the XXXI International Association of Hydrogeologists Congress "New Approaches Characterizing Groundwater Flow"*, 10-14 September, 2001. Balkema, Munich, Germany, pp. 419-423. *(.pdf)*

Timms, W., Acworth, R.I., Jankowksi, J. and Lawson, S., 2000. Groundwater quality trends related to aquitard salt storage at selected sites in the Lower Murrumbidgee alluvium, Australia. *In: O. Sililo (Editor), Proceedings of the XXX International Association of Hydrogeologists Congress "Groundwater: Past Achievements and Future Challenges".* 26 November - 1 December 2000, Cape Town, South Africa, pp. 655-660. *(.pdf)*

Timms, W., Acworth, I., Jankowski, J. and Lawson, S., 1999. Groundwater quality in a aquifer aquitard system subjected to large volume abstraction for irrigation in the Lower Murrumbidgee. *Murray-Darling Basin Groundwater Workshop "Integrated Perspec-*

tives", Griffith, pp. 131-136. *(.pdf)*

Fieldwork and software records *(.pdf)*

C.2 Hydrogeology

Bore and piezometer details *(.txt)*
Piezometer construction logs *(.pdf)*
Piezometer elevation survey *(.pdf)*
Porewater chemistry *(.pdf and .txt)*
Physical parameters *(.pdf and .txt)*
Thin section analysis of indurated clayey sand *(.pdf)*
Pump and slug test interpretation *(.pdf)*
Aquitard settlement potential *(.pdf)*

C.3 Hydrochemistry

Groundwater quality impacts *(.pdf)*
Bivariate plots of ionic ratios *(.pdf)*
Hydrochemical dataset *(.pdf and .txt)*
PHREEQC model output *(.pdf)*

C.4 Lab reports

XRD results *(.pdf)*
Stable isotope results *(.pdf)*
Radioisotope results *(.pdf)*