EVAPORATIVE DISCHARGE OF GROUNDWATER FROM THE MARGIN OF THE GREAT ARTESIAN BASIN NEAR LAKE EYRE, SOUTH AUSTRALIA

A Thesis

submitted for the degree of Doctor of Philosophy

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CONTENTS

page

ABSTRACT xiv

Declaration of Originality xvi

Acknowledgements xvii

1 INTRODUCTION 1

1.1 General Introduction and Aims 1

1.2 The Great Artesian Basin 3

1.2.1 Geography 3

1.2.1.1 general 3

1.2.1.2 project area 5

1.2.1.3 palaeogeography 12

1.2.2 Geology 16

1.2.2.1 general 16

1.2.2.2 project area 19

1.2.3 Hydrology 27

1.2.3.1 surface hydrology - general 27

1.2.3.2 surface hydrology - project area 28

1.2.3.3 hydrogeology - general 30

1.2.3.4 hydrogeology - project area 39

1.2.4 Utilization 42

1.2.4.1 general 42

1.2.4.2 project area 44

1.2.5 Importance of Diffuse Discharge 45

1.3 Stable Isotope Hydrology 47

1.4 Measurement of Evaporation from a Water Table 50

1.4.1 Introduction 50

1.4.2 Analytical and Physical Methods 53

1.4.2.1 analytical 53

1.4.2.2 hydraulic measurement and calculation 54

1.4.2.3 lysimeters 54

1.4.2.4 micrometeorological measurement 55

1.4.2.5 measurement of transpirational loss from groundwater 56

1.4.2.6 remote sensing 57

1.4.3 Tracer Methods 58

1.4.3.1 stable isotopes of water 58

1.4.3.2 chloride 59

1.4.3.3 previous studies 60

page

2 THEORY 65

2.1 Stable Isotope Relationships 65

2.1.1 General 65

2.1.2 Relationship Between Deuterium and Oxygen-18 66

2.2 Diffusion Theory 70

2.2.1 General 70

2.2.1.1 Fick's law 70

2.2.1.2 effect of companion ions 71

2.2.1.3 effect of concentration and temperature 72

2.2.1.4 measurement of diffusivity and literature values of relevance 74

2.2.2 Diffusion in Porous Media 76

2.2.2.1 geometric effects 76

2.2.2.2 anion exclusion 80

2.2.2.3 other matrix effects 84

2.2.2.4 vapour diffusion 86

2.2.2.5 liquid-vapour diffusion of water 87

2.2.2.6 non-steady state 88

2.2.3 Theory for Diffusion of Chloride and Water Isotopes in Soils 89

2.2.4 The Impedance Factor 89

2.2.4.1 impedance factor for gaseous diffusion 90

2.2.4.2 impedance factor for chloride diffusion 94

2.2.4.3 impedance factor and two phase flow for water isotopes in soil 96

2.2.4.4 modelling f_l vs θ for water isotopes 101

2.3 Convection-Diffusion Theory 106

2.4 Estimation of Vapour Flux from Suction and Chloride Data 115

3 LABORATORY TECHNIQUES AND DEVELOPMENT 118

3.1 General Soil Properties 118

3.2 Extraction of Water from Soils and Brines for Isotopic Analysis 122

3.2.1 Review of Existing Methods 122

3.2.2 Azeotropic Distillation 125

3.2.2.1 general 125

3.2.2.2 laboratory trials 129

3.2.2.3 discussion 136

page

3.2.3 Mass Balance Approach 140

3.2.3.1 general 140

3.2.3.2 laboratory trials 144

3.2.3.3 discussion 144

3.2.4 Vacuum Distillation at Room Temperature 145

3.2.4.1 laboratory trials 145

3.2.4.2 discussion 149

3.2.5 Adsorption of Vapour by Desiccants 150

3.2.5.1 laboratory trials 150

3.2.5.2 discussion 150

3.2.6 Recovery of Gypsum Crystallisation Water 152

3.2.7 Conclusion 154

3.3 Interlaboratory Comparison of Methods to Determine the Stable Isotope Composition of Soil Water 155

3.3.1 Introduction 155

3.3.2 Preparation 156

3.3.3 Results 160

3.3.4 Discussion 161

3.3.5 Conclusions 168

3.4 Electromagnetic Ground Conductivity 169

4 FIELD RESULTS 170

4.1 Field Program 170

4.1.1 Site Selection 170

4.1.2 Sampling 172

4.2 Surface Hydrology 174

4.2.1 General Conditions and Observations September 1986 to March 1988 174

4.2.2 The March 1989 Flood Event 176

4.3 Hydraulic Estimates of Leakage 181

4.4 Field Soil Profiles 184

4.4.1 Lithology and Mineralogy 184

4.4.1.1 general 184

4.4.1.2 water, gypsum, and halite 186

4.4.2 Pore Water Properties 199

4.4.2.1 chloride 200

4.4.2.2 stable isotopes 202

page
4.5 Hydrochemistry 203
4.5.1 Surface Waters 203
4.5.2 Deep Groundwaters 208
4.5.3 Shallow Groundwaters and Relationships Between Waters 212
4.5.3.1 eastern waters 212
4.5.3.2 central waters 212
4.5.3.3 western waters 217
4.5.3.4 comments 217
4.5.4 Stable Isotope Relationships 219

5 INTERPRETATION OF CHLORIDE, ISOTOPE AND SUCTION DATA 225
5.1 Hamilton Hill: Holes 4, 12, 18, 20, 21 225
5.1.1 Notes on Time Series 225
5.1.2 Evaporation Estimates 230
5.1.3 Oxygen-18 - Deuterium Relationships 236
5.1.4 Water Vapour Flux Estimates 237
5.1.5 Electromagnetic Conductivity Survey 241
5.2 Hermit Hill: Holes 1, 17, 19 243
5.2.1 Notes on Time Series 243
5.2.2 Evaporation Estimate 245
5.2.3 Oxygen-18 - Deuterium Relationships 248
5.2.4 Water Vapour Flux Estimates 248
5.2.5 Electromagnetic Conductivity Survey 248
5.3 Site GAB 6: Holes 10, 23, 24 252
5.3.1 Notes on Series 252
5.3.2 Evaporation Estimate 256
5.3.3 Oxygen-18 - Deuterium Relationships 258
5.3.4 Relative Humidity Profile 258
5.4 Other Gibber Plain sites 258
5.4.1 Evaporation and Infiltration Estimates 258
5.4.2 Oxygen-18 - Deuterium Relationships 265
5.5 Interpretation of Gypsum Water Isotope Profiles 268

6 DISCUSSION AND APPLICATION 272
6.1 Suitability of Methods 272
6.1.1 Laboratory Techniques 272
6.1.2 Convection-Diffusion Model 276

page
6.1.2.1 limitations of parameters and assumptions 276
6.2 Evaporation Estimates 281
6.2.1 Adopted Values and Uncertainties 281
6.2.2 Comparison with Other Results 283
6.2.3 Relationship Between Evaporation and Leakage 286
6.3 Areal Estimate of Leakage 287
6.4 Possible Implications for the Roxby Downs Water Supply Borefield A 290
6.5 Contribution of Great Artesian Basin Salt to Lake Eyre 292

7 SUMMARY AND CONCLUSIONS 295
7.1 Summary 295
7.1.1 General 295
7.1.2 Theoretical Development 295
7.1.3 Laboratory Technique Development 297
7.1.4 Field Results and Application 298
7.2 Scope for Further Work 302
7.3 Conclusions 305

APPENDICES
1 Lithological Description of Holes 307
2 Profile Data 320
3 Hydrochemistry 344
4 Publications resulting from the work of this thesis to February 1990 350

REFERENCES 351

LIST OF FIGURES

	page
1.2.1 Great Artesian Basin, general location	4
1.2.2 Location of project area	6
1.2.3 Climatic averages, Coober Pedy and Marree	8
1.2.4 Reconstruction of relative climate at Lakes Eyre and Frome for the last 1.5 million years	15
1.2.5 Relation of the Great Artesian Basin to adjacent basins	18
1.2.6 Correlations of hydrogeologic units in the Great Artesian Basin	20
1.2.7 Simplified hydrogeological cross-section of the Great Artesian Basin	21
1.2.8 Geological map of the project area	22
1.2.9 Structural geology of the project area with the extent of J-aquifer occurrence	25
1.2.10 Location of surface features and springs in the project area	29
1.2.11 Generalised grouping of Great Artesian Basin aquifers	31
1.2.12 Recharge areas, flow directions, and major spring groups of the Great Artesian Basin	32
1.2.13 Potentiometry of the Great Artesian Basin	35
1.2.14 Water residence times, Great Artesian Basin	36
1.2.15 Groundwater quality in the project area	40
1.2.16 Potentiometry of the artesian aquifer in the project area, before the Roxby water supply borefield installation	40
1.3.1 Craig's meteoric line of world rainfall isotopic composition	49
1.4.1 Evaporation versus groundwater depth curves of Talsma (1963)	51
2.2.1 The diffusion coefficient of sodium chloride in aqueous solutions at $25^{o}C$	73
2.2.2 Illustration of geometric effects on diffusion in a porous medium	78
2.2.3 Illustration of anisotropy of tortuosity in an oriented clay	81
2.2.4 Distribution of anions near a clay particle surface	82
2.2.5 The impedance factor for the diffusion of chloride in heavy soils as a function of water content	95
2.2.6 The effective diffusion coefficient of tritiated water in Pachappa loam at different water contents	99
2.2.7 Impedance factor as a function of water content, THO	102

page

2.2.8 Impedance factor for water isotopes as a function of water content, summary of laboratory data 102
2.2.9 Impedance factor as a function of water content modelled to match published data (i) 103
2.2.10 Impedance factor as a function of water content modelled to match published data (ii) 105
2.3.1 Calculated curves of ln $(c - c_{res})$ for different E/D rations and two different bottom boundary conditions d_1 110
2.3.2 Nomogram for the estimation of E/D from d $\ln(c - c_{res})/dz$ for shallow water tables (steady-state) 111
2.3.3 Theoretical, laboratory and field isotope profiles under steady-state evaporation 114

3.2.1 Dean-Stark apparatus for azeotropic distillation of soil water 127
3.2.2 Relative yield versus time curves for azeotropic distillation of water and soil water using toluene and kerosene 135
3.2.3 Relative yield versus time curves for azeotropic distillation of water and soil water using hexane 136
3.2.4 Temperature versus relative yield curves for azeotropic distillation of water from sand using hexane, toluene and kerosene 138
3.2.5 An example of vacuum distillation apparatus for the recovery of soil water 148
3.3.1 Results of interlaboratory comparison of methods to determine the isotopic composition of soil water, by sample type 164
3.3.2 Results of interlaboratory comparison of methods to determine the isotopic composition of soil water, by method 165
3.3.3 Results of interlaboratory comparison of methods to determine the isotopic composition of soil water, O-18 - D plot 166

4.1.1 Location of holes in the project area 171
4.2.1 Distribution of decile range numbers for March 1989 rainfall in Australia 177
4.2.2 Isohyetal map of rainfall in South Australia, March 1989 178
4.2.3 Electrical conductivity of creek water in the project area, 19-5-89 180
4.4.1-24 Water, chloride and deuterium profiles, holes 1 - 24 187-198

		page
4.5.1	Schoeller diagram, surface waters	206
4.5.2	Schoeller diagram, deep groundwaters	211
4.5.3	Schoeller diagram, eastern waters	215
4.5.4	Schoeller diagram, central (borefield) waters	216
4.5.5	Schoeller diagram, western waters	218
4.5.6	Surface and groundwaters near Lake Eyre South, O-18 -D plot	222
5.1.1	Profile comparisons, Hamilton Hill site	226
5.1.2	Hamilton Hill ln(C-Cres) vs depth plots	231
5.1.3	Hamilton Hill site pore waters, O-18 - D plot	229
5.1.4	Calculated relative humidity profile, hole 18	237
5.1.5	Calculated relative humidity profile, hole 20	238
5.1.6	Calculated relative humidity profile, hole 21	239
5.1.7	Electromagnetic conductivity survey, Hamilton Hill	242
5.2.1	Profile comparisons, Hermit Hill	244
5.2.2	Hole 1 ln(C-Cres) vs depth plot	247
5.2.3	Hermit Hill site pore waters, O-18 - D plot	247
5.2.4	Calculated relative humidity profile, hole 17	249
5.2.5	Calculated relative humidity profile, hole 19	250
5.2.6	Electromagnetic conductivity survey, Hermit Hill	242
5.3.1	Profile comparisons, GAB 6 site	253
5.3.2	Hole 10 ln(C-Cres) vs depth plot	259
5.3.3	Hole 10 pore water, O-18 - D plot	259
5.3.4	Calculated relative humidity profile, hole 24	260
5.4.1	Hole 3, ln(C-Cres) vs depth plot	261
5.4.2	Hole 11, ln(C-Cres) vs depth plot	261
5.4.3	Hole 13, ln(C-Cres) vs depth plot	261
5.4.4	Hole 14, ln(C-Cres) vs depth plot	261
5.4.5	Hole 16, ln(C-Cres) vs depth plot	266
5.4.6	Holes 2 and 3 pore waters, O-18 - D plot	266
5.4.7	Hole 11 pore waters, O-18 - D plot	267
5.5.1	Relationship between gypsum crystal and pore waters, holes 10 and 11, O-18 - D plot	269
5.5.2	Deuterium profiles of gypsum and pore waters, hole 14	271
6.2.1	Evaporation vs water table depth, project data	282
6.2.2	Evaporation vs water table depth, comparison of project and other published data	284

page

6.3.1 Physiographic units for the areal estimation of leakage in the project area 289

6.4.1 Roxby Downs water supply borefield A, observed drawdown contours mid 1989. 291

LIST OF TABLES

page

1.2.1 Monthly rainfall at Marree, 1986-1989 9

1.2.2 The geological stratigraphic column 13

1.2.3 Summary geological sequence of the project area 23

2.2.1 Diffusion coefficients of various alkali and alkaline earth metal chlorides over entire concentration ranges at 25°C 73

2.2.2 Ionic self and salt diffusion coefficients of sodium chloride, at 10°C, measured and calculated 75

2.2.3 Self diffusion coefficients of water 77

2.2.4 Partial list of published expressions relating gaseous impedance factor to air-filled and total porosity of media 92

2.2.5 Impedance factors of THO in some materials 97

2.2.6 Self-diffusion coefficients of tritiated water and chloride as a function of soil temperature in the Captina silt loam Ap horizon 100

3.1.1 Dry Bulk Density Measurements on Weathered Bulldog Shale Core 120

3.1.2 Cation Exchange Capacity of Bulldog Shale 121

3.2.1 Early azeotropic distillation verification results from the CSIRO Divisions of Soils and Water Resources,Adelaide Laboratory 130

3.2.2 Brief description of materials used in trials 132

3.2.3 Deviation from the known isotopic composition of waters azeotropically distilled with various hydrocarbons 133

3.2.4 Comparison the isotopic composition of waters extracted by azeotropic distillation with kerosene and hexane 134

3.2.5 Effect of distillation time on the isotopic composition of water azeotropically distilled with hexane, compared to a separate distillation with kerosene 137

3.2.6 Results of dilution trials to determine the isotopic composition of known pore water 146

3.2.7 Results of dilution trials to determine the isotopic composition of pore water of field samples, compared to results by azeotropic distillation with kerosene 147

3.2.8 Results of trials of room temperature vacuum distillation 149

3.2.9 Results of trials of desiccants to adsorb equilibrated water vapour at 20°C 151

	page
3.2.10 Reproducibility of azeotropic distillation with kerosene of gypsum crystallisation water	153
3.3.1 Check on reproducibility of method for preparing soil samples containing isotopically identical water	157
3.3.2 List of laboratories participating in the interlaboratory comparison of methods to detemine the isotope composition of soil water	159
3.3.3 Some physical properties of the interlaboratory test soils	160
3.3.4 Reproducibility of isotopic composition of pore water in prepared soils by a single method - azeotropic distillation	161
3.3.4 Methods employed by participating laboratories in the interlaboratory comparison of methods to extract soil water for isotopic analysis	162
3.3.5 Results summary, interlaboratory comparison of methods to determine the isotopic composition of soil water	163
4.1.1 Summary of drilling results	173
4.3.1 Hydraulic estimates of diffuse vertical leakage in the project area	183
4.5.1 Surface water chemistry	204
4.5.2 Deep groundwater chemistry (representative samples)	209
4.5.3 Shallow groundwater chemistry (representative samples)	213
4.5.4 Stable isotope analyses of surface and groundwater samples	220
5.1.1 Evaporation estimates, Hamilton Hill site	233
5.1.2 Water vapour flux estimates, Hamilton Hill site	240
5.2.1 Evaporation estimate, Hermit Hill site, hole 1	246
5.3.1 Evaporation estimate, GAB 6 site, Hole 10	257
5.4.1 Evaporation estimates, other gibber plain sites	262
5.4.2 Infiltration estimates, holes 5, 6 and 7	264
6.1.1 Summary of results from the international comparison of methods to determine the stable isotope composition of soil water, non-gypseous soils. Incomplete results to December 1989	274
6.1.2 Summary of results from the international comparison of methods to determine the stable isotope composition of soil water, gypseous soils. Incomplete results to December 1989	275
6.2.1 Adopted evaporation estimates	283
6.3.1 Physiographic units and assigned deep aquifer leakage rates	288

LIST OF PLATES

	page
1.1 General view of gibber plain near Lake Eyre South.	10
1.2 Reed (*Phragmites*) covered mound spring near Bopeechie.	10
4.1 Investigator drilling rig on bare gibber plain at hole 8.	175
4.2 Site GAB 6, looking from hole 23 on a slight rise to hole 24 in a slight depression, with vegetation beyond	175
4.3 Screechowl Creek flowing with saline water toward Lake Eyre	179
5.1 View looking south along the electromagnetic conductivity transect line past the site of holes 1, 17 and 19 (at vehicle) and some mound springs to Hermit Hill.	251

ABSTRACT

The Great Artesian Basin is Australia's largest groundwater basin and represents an important resource. Its overall water budget implies that about half the discharge is by bores, and of the remaining discharge only a small portion emerges in springs, the rest is assumed lost by leakage to the overlying water table, from where it evaporates.

This leakage is assessed using a steady-state convection-diffusion model to estimate evaporation from vertical profiles of the concentrations of chloride and deuterium in the unsaturated and saturated zones. The theory has been extended to account for shallow water tables, and a nomogram constructed that enables the estimation of evaporation from a plot of $\ln(c-c_{res})$ vs depth, as is used with an infinite bottom boundary condition here and in earlier work. An important parameter needed to make the calculation is the impedance factor; this is examined on detail using existing literature.

To make the estimates of evaporation, isotopically representative water must be extracted from soil samples. Extraction methods have been extensively investigated. Azeotropic distillation with hydrocarbons is a robust and reliable technique. The use of toluene gives best results for water, and kerosene for non-gypseous porous materials, with accuracies of about 1.5 and 0.2‰ for δD and $\delta^{18}O$ respectively. Errors increase at low soil water contents (high matric suction). The use of hexane, with distillation time restricted to two to three hours, provides pore water from gypseous soils. The distillate is isotopically biased by about -3‰ and -1.1‰ for δD and $\delta^{18}O$ respectively, with an accuracy of about 2‰ and 0.3‰. This analytical advance permits the extraction of both bore and crystal water from the same gypseous sample.

An international intercomparison of techniques to determine the

isotopic composition of soil water was conducted. This reveals a much greater scatter of results than the accuracy claimed by most workers. Comparisons of the absolute values of soil isotope data between laboratories using different techniques should be made with caution.

Twenty-four holes were drilled in a field program near Lake Eyre in South Australia. Evaporation estimates could be made from 11 holes, and range from about 0.5 to 4.5 mm yr^{-1}. The average leakage is estimated to be between 2 and 4 mm yr^{-1} above the artesian parts of the basin margin in the field area, and amounts to five to ten times the combined discharge of springs in the area. The measurements show that leakage to the water table and subsequent evaporation can account for a large part of the water balance of the basin, and this conclusion is also supported by other isotopic and hydrochemical evidence.

The natural leakage of artesian water from the basin margin gives a limit to the amount of the resource that can safely be withdrawn by users, in this case a large mining venture. The mine has an expected lifetime of several decades and careful management of the borefield is required to sustain the water supply in the long term.

DECLARATION OF ORIGINALITY

I certify that this thesis does not incorporate without acknowledgement any material previously submitted for a degree or diploma in any university; and to the best of my knowledge and belief it does not contain any material previously published or written by any other person except where due reference is made in the text.

Peter H. Woods

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CHAPTER 1

INTRODUCTION

1.1 GENERAL INTRODUCTION AND AIMS

Discharge by evaporation of groundwater from shallow water tables is an important component of the water balance of many aquifer systems in the drier parts of the world. The magnitude of the discharge by this means has generally been estimated by difference, the "left-over" in a general water balance. Only in recent years has it been possible to directly estimate evaporation from shallow water tables in arid groundwater discharge zones, and relate this to an overall groundwater basin water balance (e.g. Menenti, 1984, Allison & Barnes, 1985, Christmann & Sonntag, 1987). Many of these studies, such as the last two cited, use the vertical concentration profiles of solutes in the soil, namely the stable isotopes deuterium and oxygen-18 in water itself, or the anion chloride.

The Great Artesian Basin (GAB) is Australia's largest groundwater basin, and large by world standards, occupying some 1.7 million square kilometres. Most of the Basin's groundwater flow is entirely internal, and its water balance indicates that a large proportion of its natural discharge is by leakage through confining beds to the water table near the surface, and subsequent loss by evaporation. The Basin provides the only reliable water supply to pastoral and extractive industries (hydrocarbons and mining) over large parts of the states of Queensland, New South Wales and South Australia, and a small part of the Northern Territory.

A major new user of the GAB groundwater is the Olympic Dam Project, a copper-uranium-gold mine at Roxby Downs in South Australia. It draws the majority of its water requirements from the margin of the GAB near Lake Eyre South. This margin is part of the main natural discharge area of the Basin. The amount of water that can be safely withdrawn from the borefield without adverse effects on water quality is closely related to the amount formerly lost from the aquifer by leakage and evaporation from the shallow water table beneath arid stony plains, and from springs and their surrounds. While spring discharge is visible and important, the water balance suggests that most of the loss is by diffuse discharge at low rates over a large area, where the water evaporates unseen from the water table a few metres below the ground surface.

This study aims to estimate evaporation from shallow water tables within the Olympic Dam Project borefield and adjacent areas, and relate this to the water budget of the GAB aquifer in this strip of the Basin margin. Convection - diffusion theory is applied to estimate the upward flux of water from a number of field profiles, and these point estimates are extended by physiographic association to provide an overall estimate of leakage from the aquifer in the project area. This can be related to the likely safe yield of the area for sustainable commercial exploitation.

In order to use the convection-diffusion approach to estimate evaporation of groundwater, the assumptions and uncertainty of the parameters involved need evaluation. In the past the presence of gypsum in soil meant that such samples could not be analysed for the stable isotope composition of pore water, as the water of crystallisation was also removed by the available extraction methods; evaporation estimates could be made on incomplete profile data or not at all. An early requirement of the project was to overcome this technical difficultly, as the majority of field profiles contained abundant gypsum. Solving this problem enables the comparison of the isotope composition of water of crystallisation in gypsum to that of soil pore water in the same sample, and so shed some light on the origin of the gypsum.

1.2 THE GREAT ARTESIAN BASIN

1.2.1 GEOGRAPHY

1.2.1.1 General

The Great Artesian Basin is an enormous geological structure that occupies 1.7 x 10^6 km^2, or about one-fifth of Australia, extending beneath parts of Queensland, the Northern Territory, New South Wales, and South Australia (Fig. 1.2.1). Underlying mostly arid and semi-arid regions where surface water supplies are sparse and unreliable, its subterranean water is a valuable resource (Habermehl, 1980). From prehistory its springs were used by fauna and Aboriginal peoples, the latter well evidenced by abundant flakes from stone tool making still seen at some springs. The discovery of the basin's groundwater supplies around 1880 by white settlers, and subsequent development, allowed vast areas to be opened up to grazing, and provided town water supplies (ibid.), a role still played over a century later. The petroleum and natural gas industry at Moomba and surrounding fields, and the large mining development at Roxby Downs in South Australia have been more recent users on a larger scale than previous individual enterprises. The establishment of the Roxby Downs borefield in the 1980s was a large motivator of the field work of this thesis. To put this project in

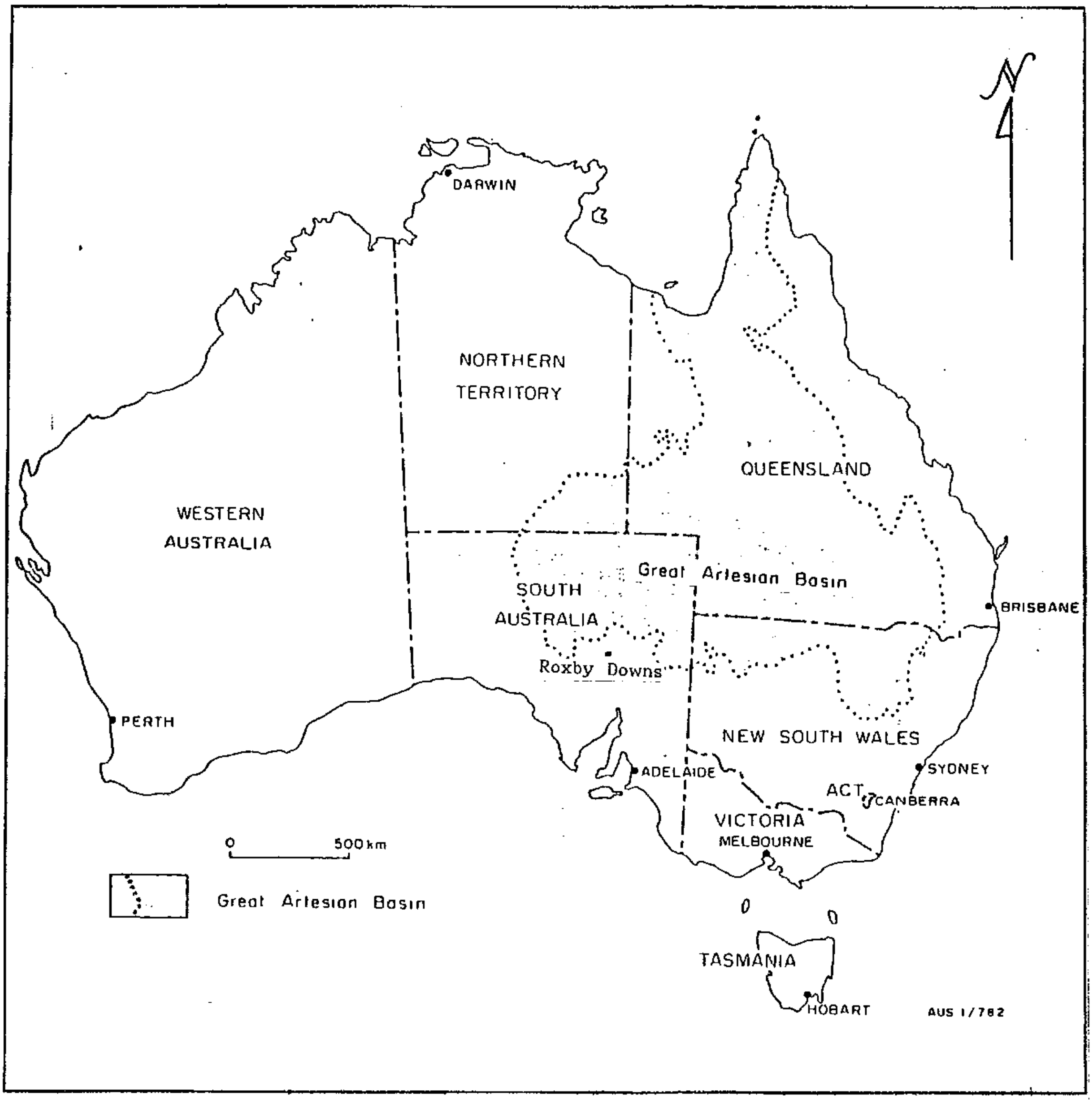


FIGURE 1.2.1 Great Artesian Basin, general location

context, the geology, hydrology and use of the Great Artesian Basin (GAB) and in particular the Roxby Downs water supply area are now considered.

1.2.1.2 Project Area

The extended project area is shown in figure 1.2.2 and stretches from Coober Pedy in the west (longitude 134° 30' E) to Lake Eyre South in the east (longitude 137° 30' E), at a latitude of 28° 45' to 29° 30' S. Most work was carried out within about 30 km of Lake Eyre South, which will be called simply the project area.

The extended project area is reasonably flat, and goes to the shores of Lake Eyre which encloses Australia's lowest topography of -14.8m AHD (Australian Height Datum, which is virtually mean sea level: Dulhunty, 1987). To the south east the area is bounded by the Willouran Ranges, at the north-west end of the Flinders Ranges, and to the north-east by Lake Eyre. To the southwest and west are the low Stuart Ranges that run through Coober Pedy, a series of breakaways where an earlier plain is being eroded into Lake Eyre to form a new plain (Jessup & Norris, 1971). The new plain commonly has an armouring of siliceous pebbles called gibbers, thought to be remnants of the siliceous capping (silcrete) formed during the Tertiary and still existing on the old plain level (ibid.; Plate 1.1). Lake Cadibarrawirracanna occupies a local depression in the west between William Creek and Coober Pedy, and is separated from Lake Eyre by the low Peake and Dennison Ranges. Some parts around Lake Cadibarrawirracanna contain longitudinal sand dunes a few metres high. Many of these features are apparent on the geological map, figure 1.2.8, discussed shortly.

The climate of the area is very arid by Australian standards, described as a very hot dry desert climate with a short cool to cold winter, Köppen's BWh classification (Laut *et al.*, 1977). The mean and

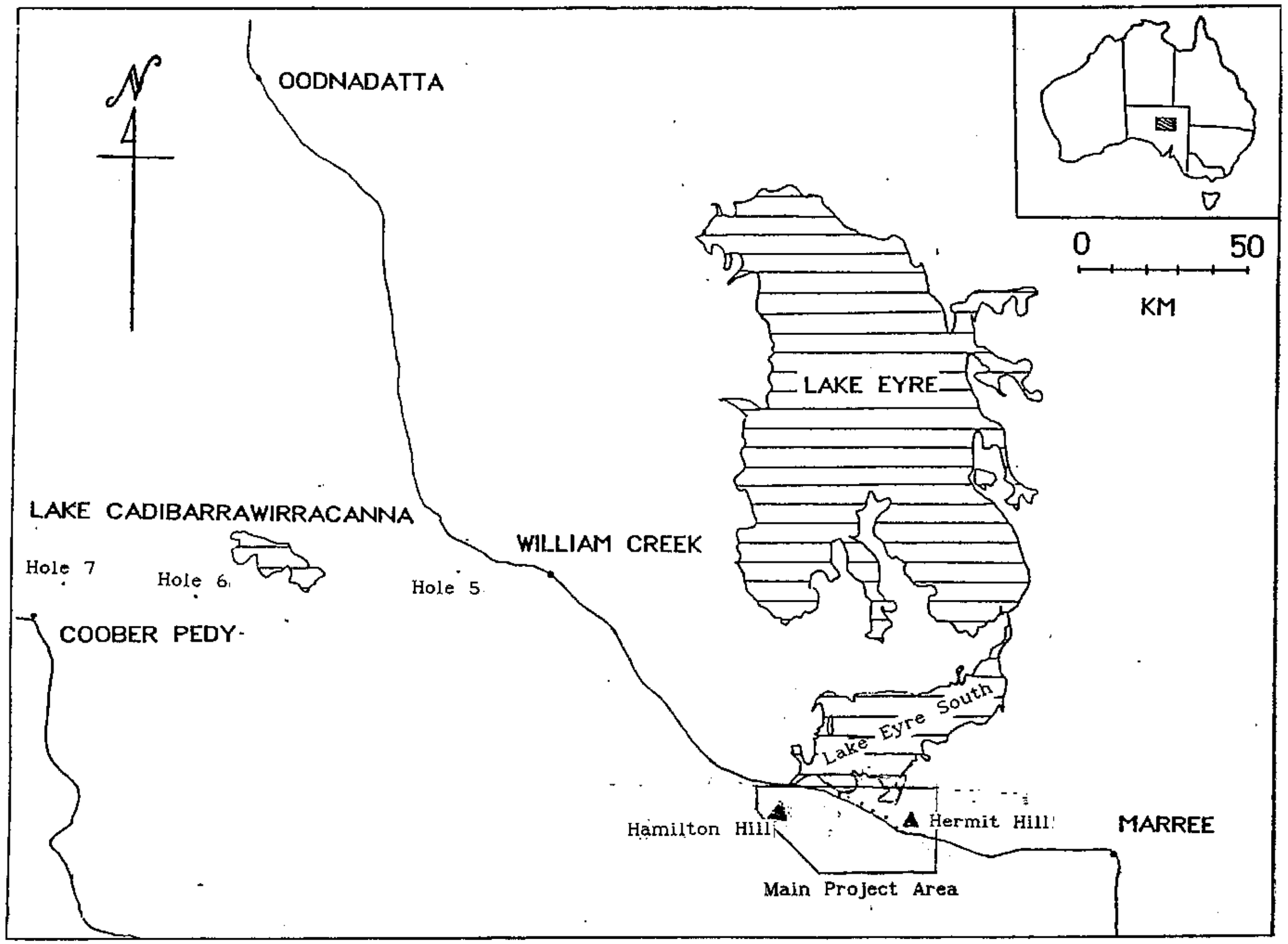
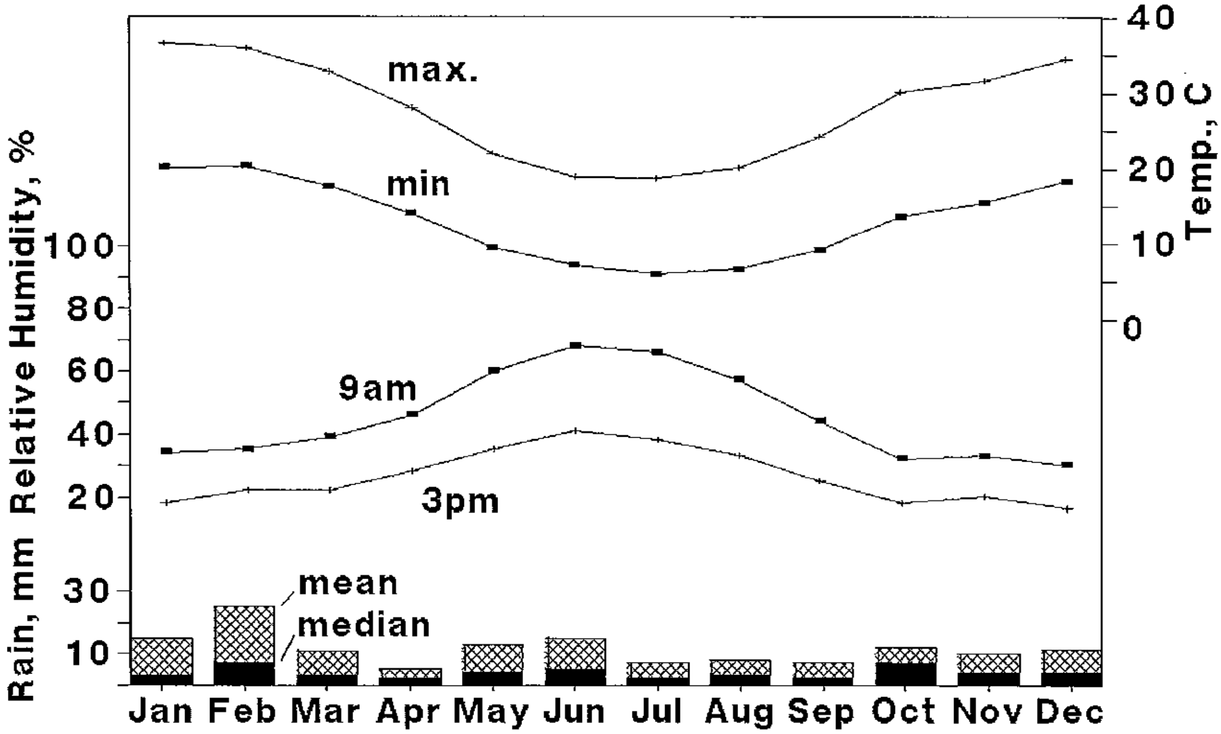


FIGURE 1.2.2 Location of Project Area

median monthly rainfall, monthly average 9am and 3pm humidity, and maximum and minimum temperatures for Marree and Coober Pedy are shown in figure 1.2.3. Summers are hot and dry, winters cool and dry. The annual average rainfall is about 125 to 150 mm, but falls are extremely irregular, and may occur at any time of year, though some claim a February maximum (Tetzlaff & Bye, 1978). Winter rainfall from the south is slightly more reliable, but the highest individual falls tend to be in the hot months when tropical moisture feeds in from the north-west to north-east on rare occasions. Monthly rainfall at Marree over the project time (1986 – 1989) is shown (table 1.2.1) and was generally above average, extremely so in March 1989. Average annual potential evaporation is about 3000 mm, and lake evaporation 2000 mm a^{-1} (Tetzlaff & Bye, 1978) when free water surfaces are large.

The soils of the extended project area are first described in some detail by Jessup (1960a,b), and are discussed further by Laut *et al.* (1977). Of Jessup's seven soil regions, those represented in the extended project area are lateritic soils (on the Stuart Ranges near Coober Pedy), stony tableland soils (gibber plains), and sandridges between Coober Pedy and William Creek. The soils and shallow geology have been interpreted as due to cycles of relatively dry and humid periods (Jessup & Norris, 1971; Wells & Callen, 1986). Soils tend to be sodic and saline, and gypsum is a common constituent and may form massive gypsites or occur at shallow depth in soil profiles in veins or as a cement. Very similar soils to the project area (stony tableland soils), developed on the same Cretaceous shale parent material, near Andamooka some 110 km south of the project area, are described by Milnes *et al.* (1987, p16), with mineralogical data. The soil is armoured with siliceous pebbles. The top 1 cm is a vesicular crust of low clay content (15%), underlain by 44-49% clay material, with halite (NaCl) and calcite

Coober Pedy



Marree Post Office

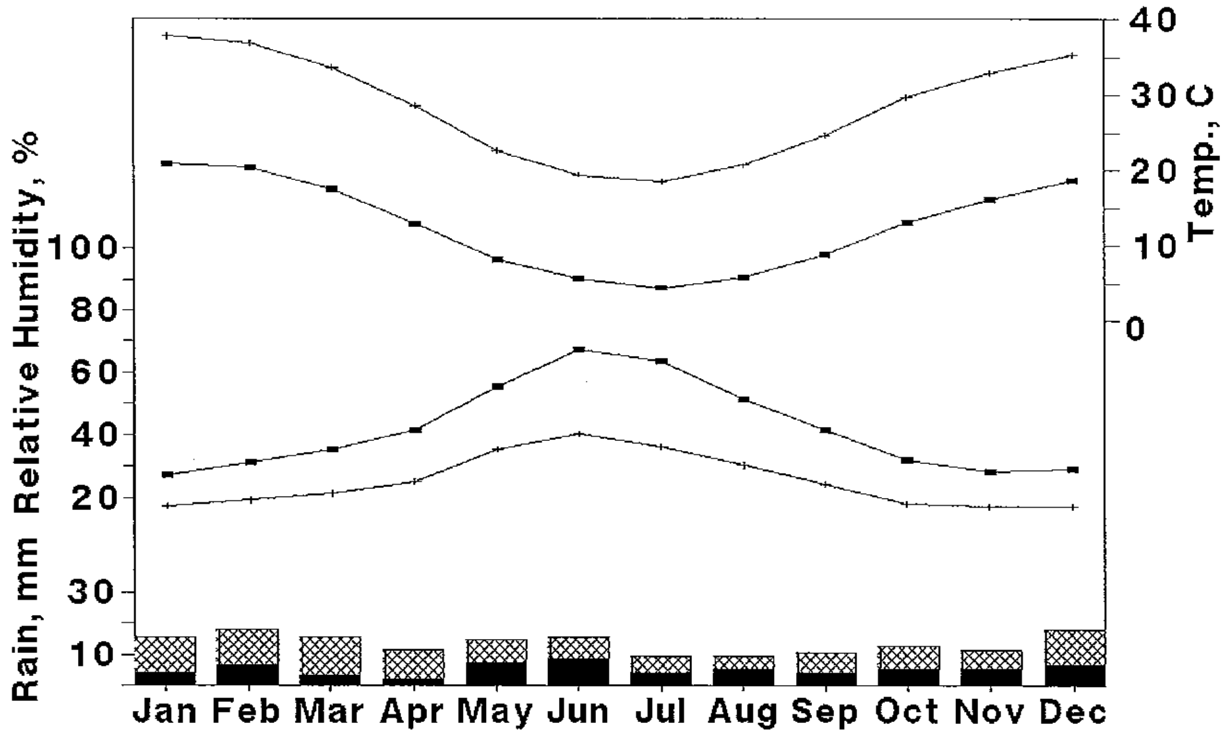


FIGURE 1.2.3 Climatic Averages, Coober Pedy and Marree

TABLE 1.2.1

Monthly Rainfall at Marree, 1985-1989

Month	1985	1986	1987	1988	1989
January	0	0	24	0	1
February	2	5	43	0	0
March	1	0	31	32	180
April	0	0	0	1	2
May	13	49	2	52	7
June	2	2	10	11	21
July	7	25	2	13	20
August	17	33	19	6	2
September	0	7	2	2	0
October	26	44	5	0	21
November	45	11	1	0	30
December	8	6	16	81	
Totals	121	181	155	186	

Notes: Data is rounded to nearest millimetre. Courtesy Australian Bureau of Meteorology.

PLATE 1.1 General view of gibber plain near Lake Eyre South.

October 1986

PLATE 1.2 Reed (*Phragmites*) covered mound spring in the project area.

Bopeechie Springs, July 1987

in small quantities, and gypsum appearing at 30-40 cm and increasing down the profile.

The vegetation of the area reflects the low rainfall, high temperatures and potential evaporation, and generally poor soils. It is desertic except for some mound springs and bores, and major creekbeds, and is described in Laut *et al.* (1977) and by Symon (1985a,b). Much of the gibber plain of the extended field area is bare except for ephemeral grasses and herbs in wet seasons. Some scattered drought-hardy small shrubs and tussock grasses occur, especially on slight rises and some sand dunes. Trees are rare, those observed by the author include red mallee (*Eucalyptus socialis*), native apricot (*Pittosporum phylliraeoides*), and some acacias (e.g. *A. stenophylla*, river-cooba or eumong). Greater variety and size of vegetation is observed lining the larger creek beds, including larger acacias and eucalypts. For more detail and fuller species lists readers are referred to Symon (1985a,b).

The vegetation associated with mound springs has attracted a lot of attention since the announcement of the Roxby Downs development, with two major initial surveys (Symon (a,b) and Alcock in Greenslade *et al.*, 1985, and Kinhill Stearns, 1984), and ongoing monitoring by T. Fatchen and Associates for Roxby Management Services, as well as the S.A. Department of Conservation and Lands. The most common spring associated types are sedges (notably *Cyperus gymnocaulos* and *C. laevegatus*), the tall reed *Phragmites australis*, and bulrushes (*Typha* sp.) (Symon 1985a, Kinhill Stearns, 1984), which also occur in other Australian wetlands. Some plants appear confined to GAB springs, notably a rare button grass *Eriocaulon carsoni*, confirmed in the project area (ibid.). Many of the plants found on and around the springs are heavily grazed by cattle if accessible. A small reed-covered mound spring is shown in Plate 1.2.

1.2.1.3 Palaeogeography

Enough work has been published in the last two decades to enable a reconstruction to be made of the palaeogeography of the Lake Eyre region since the Cretaceous seas retreated. A stratigraphic column of the names and order of eras, periods and epochs, and their radiometric ages, is given in table 1.2.2. The inferred climates from the Tertiary to the Quaternary given below are summarized from Wells and Callen (1986) unless otherwise indicated.

During the Palaeocene (from 65 Ma ago), as Australia and Antartica were separating, the fluvial sediments of the Eyre Formation were being deposited in the Lake Eyre Basin, and the climate was quite humid and cool, supporting rainforest similar to that found in western Tasmania today. By the late Eocene (ended 38.5 Ma ago) the climate had become drier, with less fluvial activity, and intensive weathering leading to the development of silcrete and laterite. In the Oligocene and Early Miocene the first *Eucalyptus* and related genus forests appeared. During this time minor earth movements (tectonism) cut off surface drainage to the sea, and the first of a series of large lakes formed, which have remnants to the present day. During all this time, the Australian continent was drifting towards the equator and experiencing warmer and drier conditions. The Etadunna Formation formed in the lake bed ancestral to Lake Eyre (sometimes called Lake Dieri, Dulhunty, 1983), and perhaps ancestral Lake Frome. Grasslands appeared, but some rivers were still perennial and supported forests, and there is evidence of abundant fauna. The climate began to alternate between drier and more humid phases, gradually becoming drier overall. There is evidence of seasonal aridity, with gypsum deposited by evaporating groundwaters (as today), less forest, and more grassland and shrubland during a dry part of the

STRATIGRAPHIC COLUMN

ERA	PERIOD	EPOCH		AGE OF BASE (in millions of years before present-Ma)
	Quaternary	Holocene		10 000 years
		Pleistocene		1·8 Ma
Cainozoic	Tertiary	Pliocene	Neogene	5
		Miocene		23·5
		Oligocene	Palaeogene	38·5
		Eocene		53·5
		Palaeocene		65
Mesozoic	Cretaceous			135
	Jurassic			190
	Triassic			225
Palaeozoic	Permian			280
	Carboniferous			345
	Devonian			395
	Silurian			430
	Ordovician			500
	Cambrian			570
Proterozoic	Adelaidean	Precambrian		uncertain; ?800-1 100
	Carpentarian			1 800
	Early Proterozoic			2 500
Archaean				

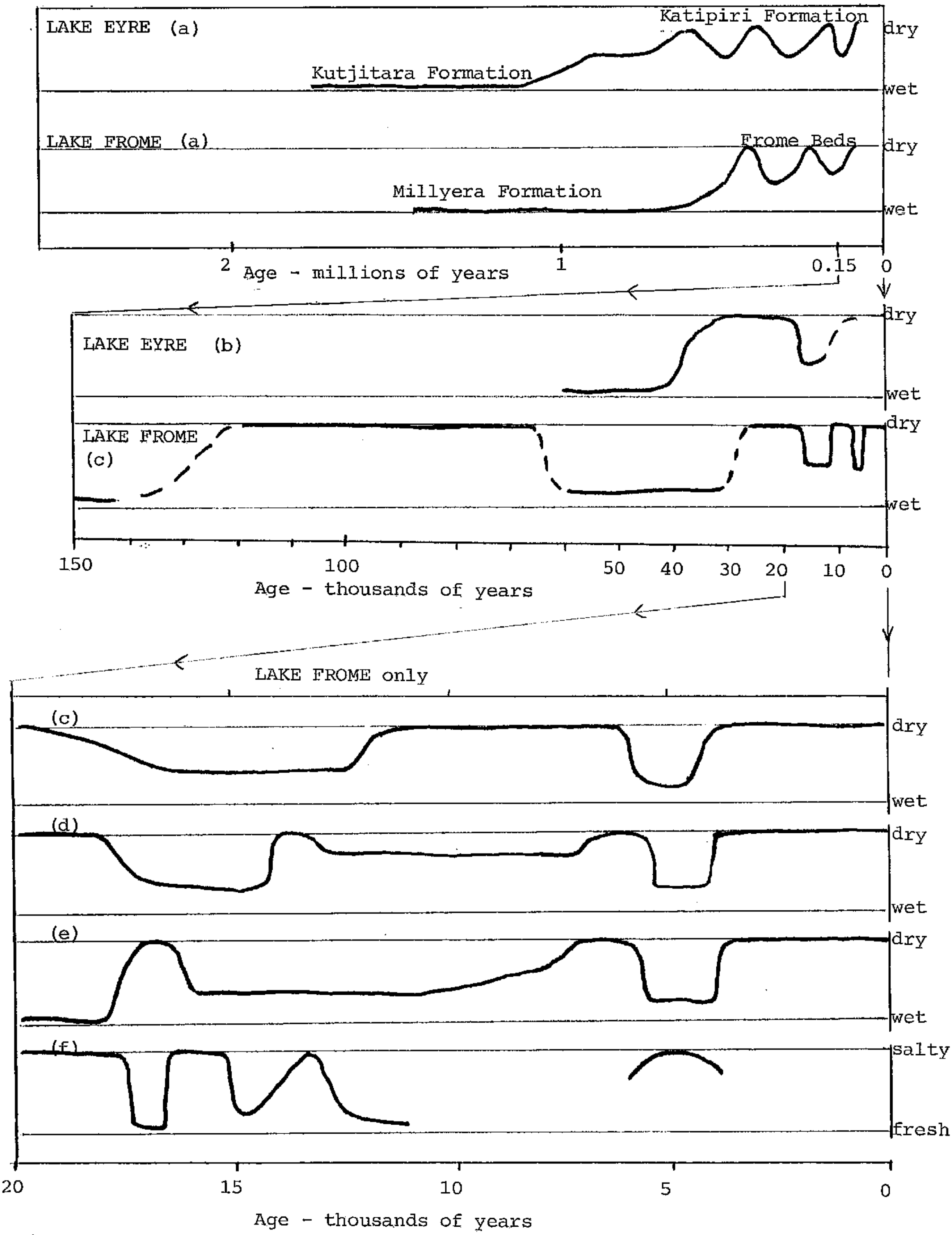
Formation of Earth's crust about 4 600 million years ago

TABLE 1.2.2 The Geological Stratigraphic Column.
from Ludbrook, 1980.

Pliocene. Events of the late Pliocene and Pleistocene developed in parallel to the global glacial climatic changes, with increased stream flow and lake-full conditions in early glacial cycles, succeeded by desiccation and dune building, and deposition of evaporites during glacial maxima. The overall trend was still one of increasing aridity, with the last most arid phase about 18 000 years BP (Before Present).Wasson & Clark (1988) have compiled the data of Wells & Callen (1986) for Lakes Eyre and Frome for the last one or two million years (Fig. 1.2.4).

The two lakes are far enough apart to show some differences in high and low water level events. More information is available for Lake Frome however, so is presented in the rest of figure 1.2.4 as a guide for Lake Eyre. Work on Lake Eyre by Magee *et al.*. (1988) and Lake Frome (Bowler & Magee, 1988) enable greater detail to be given for the last 150 ka. Both works are based on stratigraphic interpretation, with some carbon-14 dates for the latter. Lake Frome was dry from about 120 ka to 65 ka, and both lakes wet from then, Lake Eyre drying out about 40 ka and Lake Frome about 30 ka. Both returned to an intermediate lacustrine stage from 16.5 ka to 12 ka BP, when the record ends for Lake Eyre. Lake Frome had a final intermediate lacustrine period from 6 to 4.5 ka BP, which could have affected Lake Eyre as well, before returning to the playas evident today.

For the last 20 ka BP, other reconstructions for Lake Frome are also available. Singh and Luly's (1988) information is based on stratigraphic and pollen analysis of a core, and a history of the area's vegetation can be inferred (not presented here). Their reconstruction agrees fairly well with that of Bowler and Magee (1988), except for the end of the lacustrine period beginning about 17 ka BP, determined as at about 14 ka by Singh and Luly, and at 12 ka by Bowler and Maggee. An earlier result of Bowler *et al.* (1986) based on stratigraphy and C dating is also shown, it does not correlate well with the later result prior to 7 ka BP. The



Sources:
(a) Wasson & Clark, 1988. (b) Magee et al., 1988. (c) Bowler & Magee, 1988.
(d) Singh & Luly, 1988. (e) Bowler et al., 1986. (f) Ullman & McCloud, 1984.

FIGURE 1.2.4 Reconstruction of relative climate at Lakes Eyre and Frome for the last 1.5 million years *salinity* of the lake water (rather than water level alone) is inferred from the sodium content of gypsum precipitated at the time (Ullman & McCloud, 1984). Salinity apparently oscillated from 17 ka to 12 ka BP, and the 6 - 4 ka filling was quite salty, perhaps due to the presence of evaporites which accumulated in the preceding dry millennia.

Some of these wet and dry periods have occurred since the appearance of man in Australia 30, 40 or perhaps more millennia ago. The climate found today however has probably been similar for the last 4000 years.

A little can be deduced about the palaeohydrology of the GAB. The reasonable correspondence of ^{36}Cl and hydrodynamic ages of water show the flow paths and rates have been virtually the same for at least the last million years, and probably much longer, extending back into the late Tertiary. Large deposits from now-extinct springs above the current piezometric surface are evident at Hamilton Hill (Habermehl, 1980, Thompson & Barnett, 1985), Dalhousie (Krieg, 1986, Williams & Holmes, 1978), and other sites, dated tentatively to the Pleistocene. They were formed at a time when both piezometric head and the general plain level were higher than at present. When more recent erosion lowered the general landscape level, new springs at lower elevations dissipated the head and lead to the drying up of the higher, older springs.

1.2.2 GEOLOGY

1.2.2.1 General

Setting

The main aquifers of the GAB consist of sandstones interbedded with shales, dated to the Lower Jurassic to Upper Cretaceous, or roughly 180 to 65 million years old. Over the largest part of the basin, including the project area, they are assigned to the Eromanga Basin (Senior *et al.*, 1978), but assigned to the Carpentaria Basin north of the Euroka Ridge, and the Surat Basin east of the Nebine Ridge (Power & Devine, 1970, Exon & Senior, 1976). There appears to be a subsurface geological connection to the Cretaceous Monash Formation beneath the younger (Tertiary) Murray Basin in the south, via the Darling Corridor (Thornton, 1972, Power & Devine, 1970). These relationships are shown in figure 1.2.5. The sediments reach a maximum thickness of about 3000m. The Eromanga Basin is in part underlain by older sedimentary basins such as the Adavale, Drummond, Georgina, Galilee and Cooper (Senior *et al.*, 1978), some of which are of interest for hydrocarbons, as is the Eromanga itself (Sprigg, 1986, Armstrong & Barr, 1986). Other parts are underlain by crystalline bedrock, or very old sediments such as the Adelaide Geosyncline (in the project area on the south west margin of the basin). The sediments of the Eromanga Basin outcrop or subcrop beneath shallow recent sediments along its margins. Towards the centre of the GAB the Eromanga is overlain by the Cainozoic Lake Eyre and Callabonna Basins (Wells & Callen, 1986), which reach up to 150m thickness.

Stratigraphy and Depositional Environment

The stratigraphy of the Eromanga Basin has been fairly well investigated despite its vast extent, spurred on greatly by petroleum exploration. Recently published information appears in the volume edited by Gravestock *et al.* (1986); more detailed information is held for some parts of the basin by petroleum companies, and the stratigraphy of the area undergoes continuing reappraisal. Different unit names have often been assigned in the different states, the correlations and basic

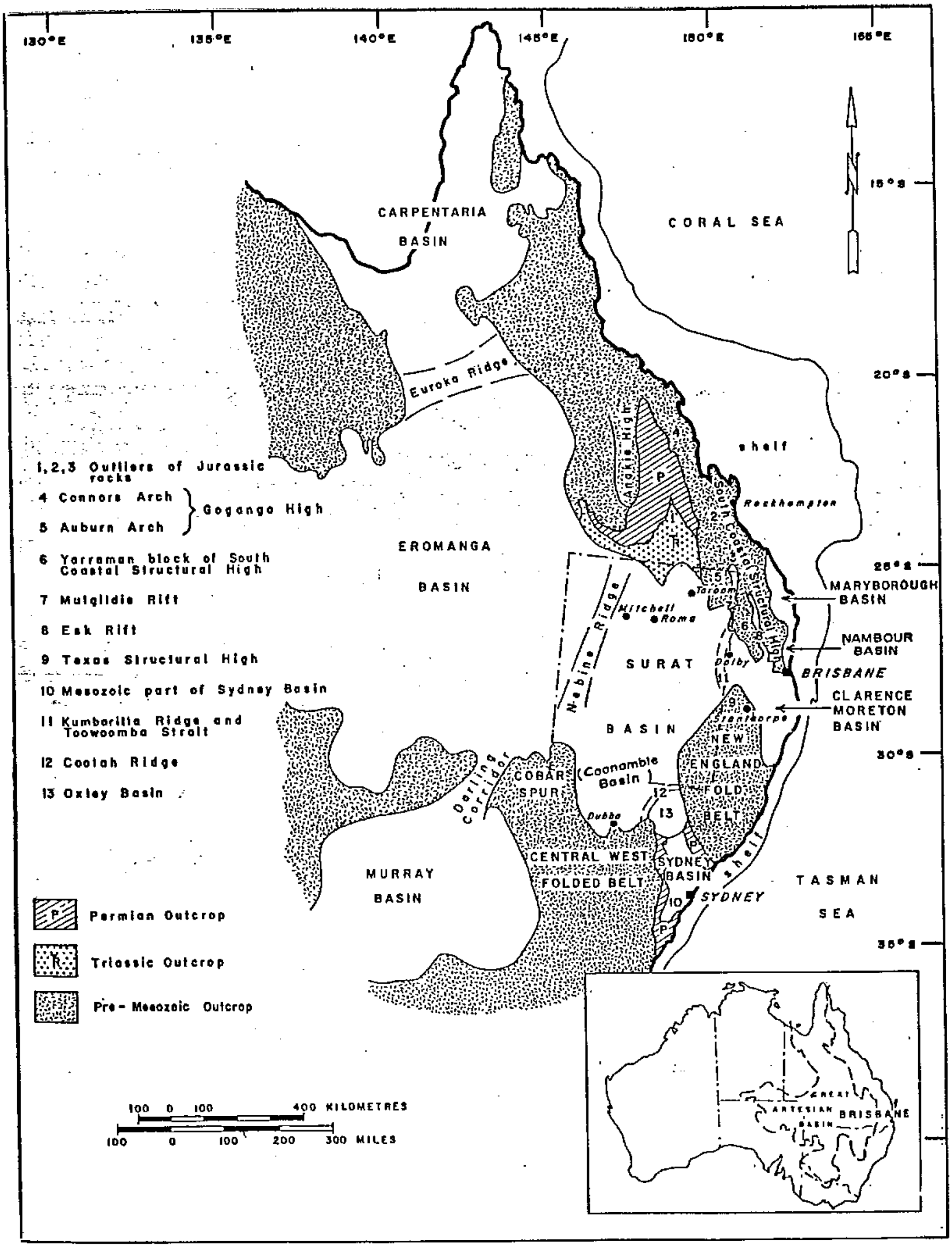


FIGURE 1.2.5 Relation of the Great Artesian Basin to Adjacent Basins (Power & Devine, 1970)

stratigraphy are shown in figure 1.2.6 (Habermehl, 1980). The Mesozoic sediments form a shallow dish shape in cross-section, with some local interruptions by faulting and gentle folding (Fig. 1.2.7).

The sediments themselves consist of interbedded sandstones, mudstones and siltstones (the latter two often referred to collectively as shales), laid down in continental and marine environments as shallow seas repeatedly invaded (transgressed) and retreated (regressed) from the continent (see Frakes *et al.*, 1987). Much of the sandstone of the late Jurassic appears to have been laid down in fluviatile continental environments (Nugent, 1969, Wopfner *et al.*, 1970), with some fine grained lacustrine beds (Nugent, 1969). The early Cretaceous sediments show that shallow marine conditions then dominated, leading to mainly mudstones and shales (Freytag, 1966, Wopfner *et al.*, 1970), though there were apparently two transgressive-regressive cycles over the basin as a whole, over about 20 million years (Exon & Senior, 1976, Frakes *et al.*, 1987). The overlying Tertiary and younger sediments are terrestrial, of mainly fluvial and lacustrine origin (Wells & Callen, 1986).

1.2.2.2 Project Area

The geology of the margin of the Eromanga Basin in South Australia is discussed by Forbes (1986). The geologic sequence of the project area is given in table 1.2.3 (A.G.C., 1984), and the outcrop pattern in a geological map (Fig. 1.2.8).

The oldest rocks are slightly metamorphosed, folded sediments of the Adelaide Geosyncline that outcrop in the Willouran Ranges in the south-east (and much of the Flinders Ranges beyond), and as inliers, notably Hermit Hill in the project area and the Peake and Denison Ranges west of Lake Eyre North. Other basement rocks may be present in the

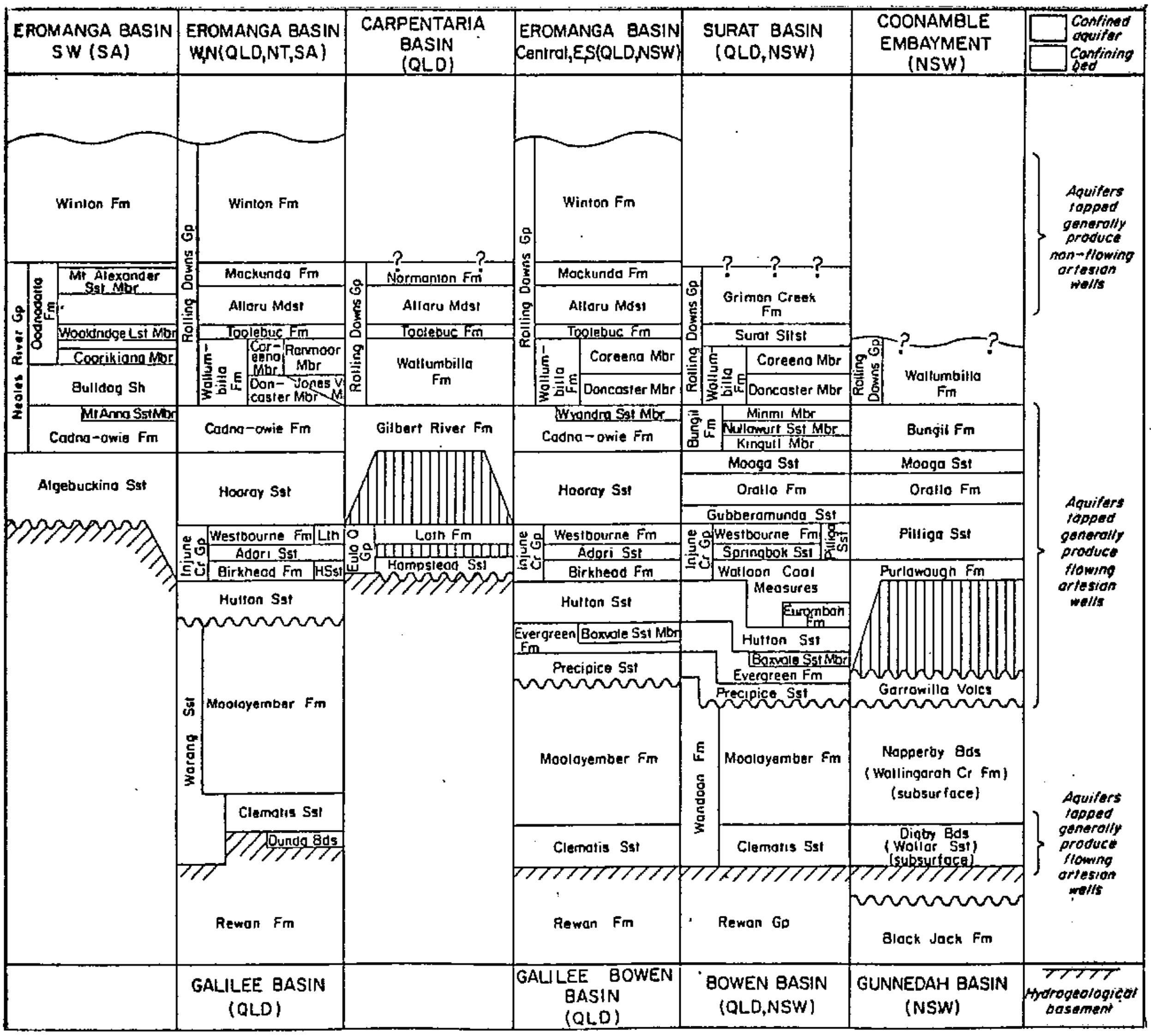


FIGURE 1.2.6 Correlation of Hydrogeological Units in the Great Artesian Basin (Habermehl, 1980).

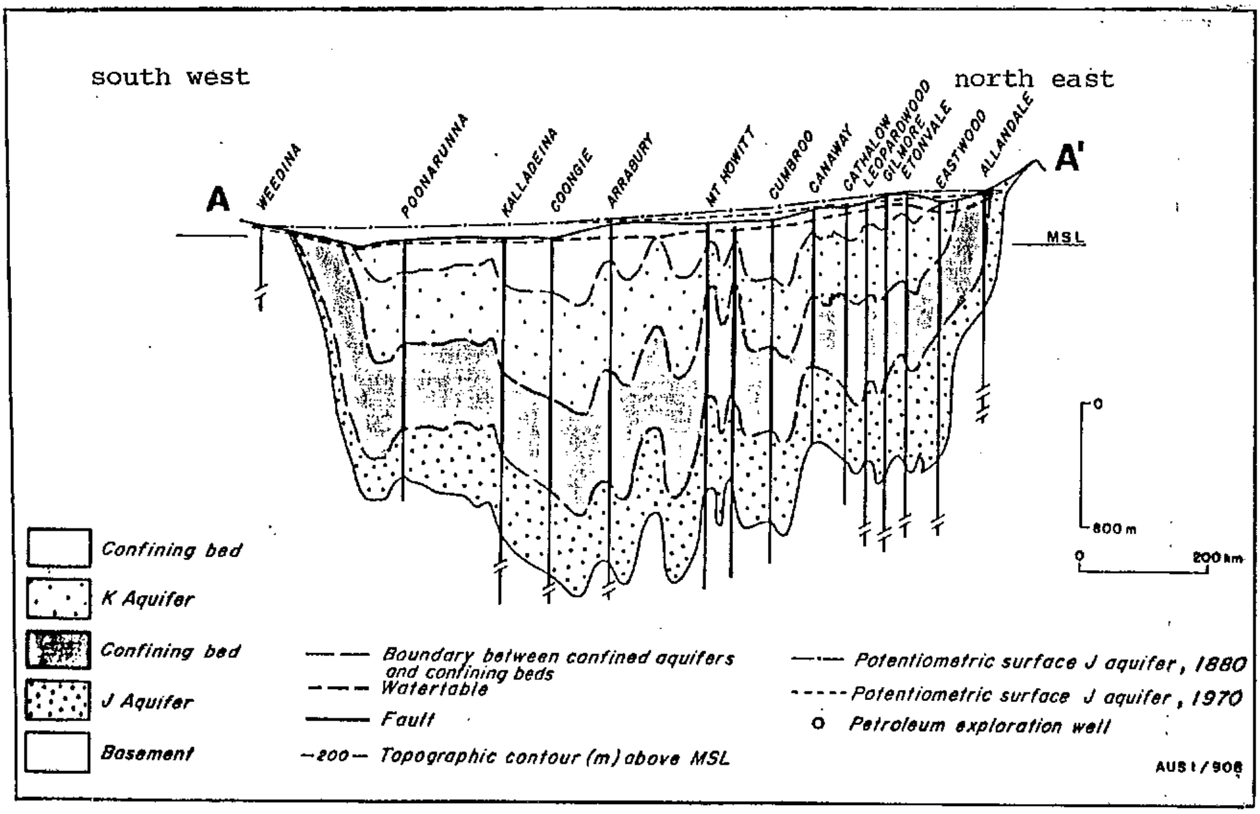


FIGURE 1.2.7 Simplified Hydrogeological Cross-section of the Great Artesian Basin (Habermehl, 1980).

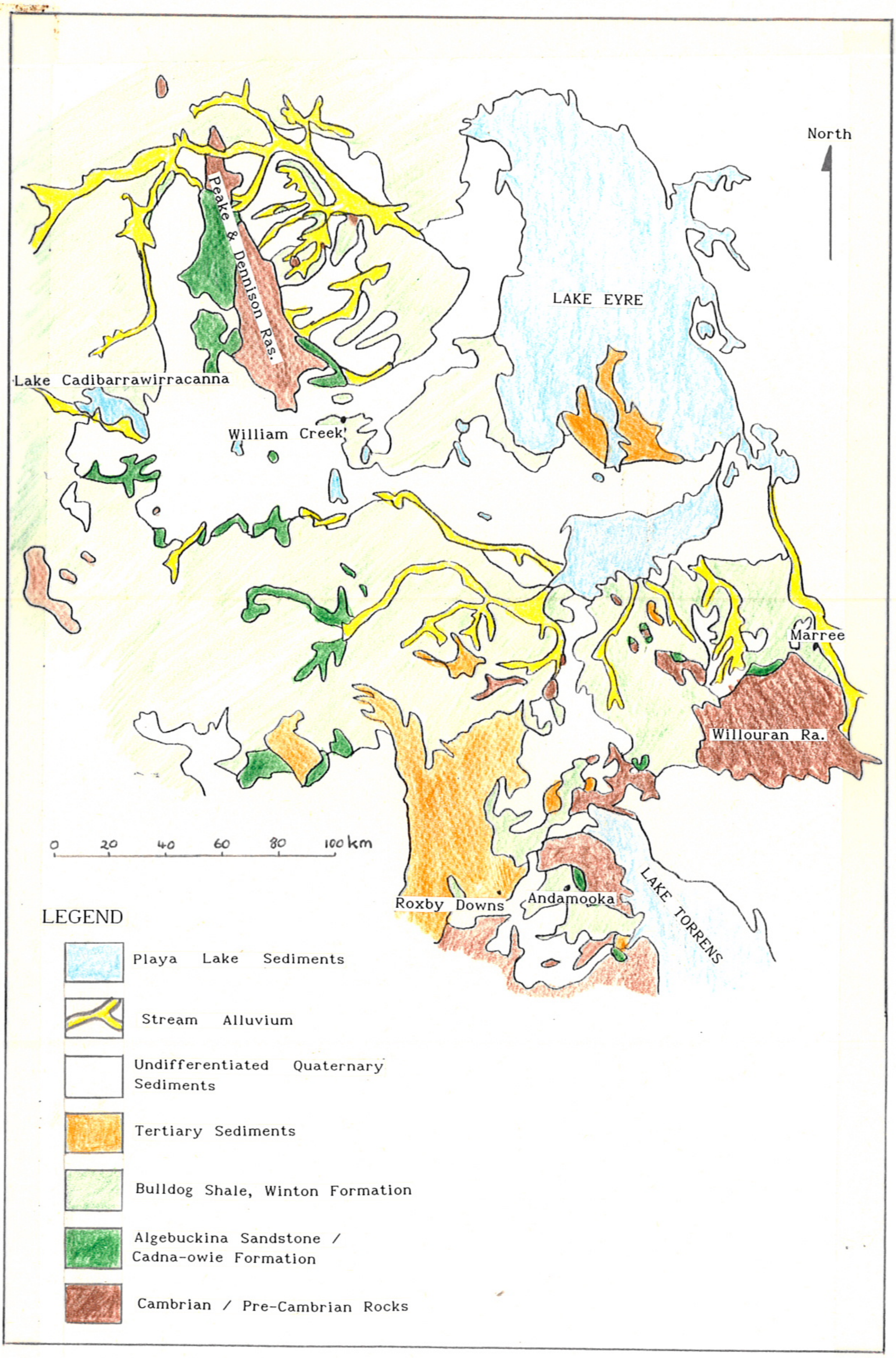


FIGURE 1.2.8 Geological Map of the Project Area.

after The Geological Survey of South Australia.

TABLE 1.2.3

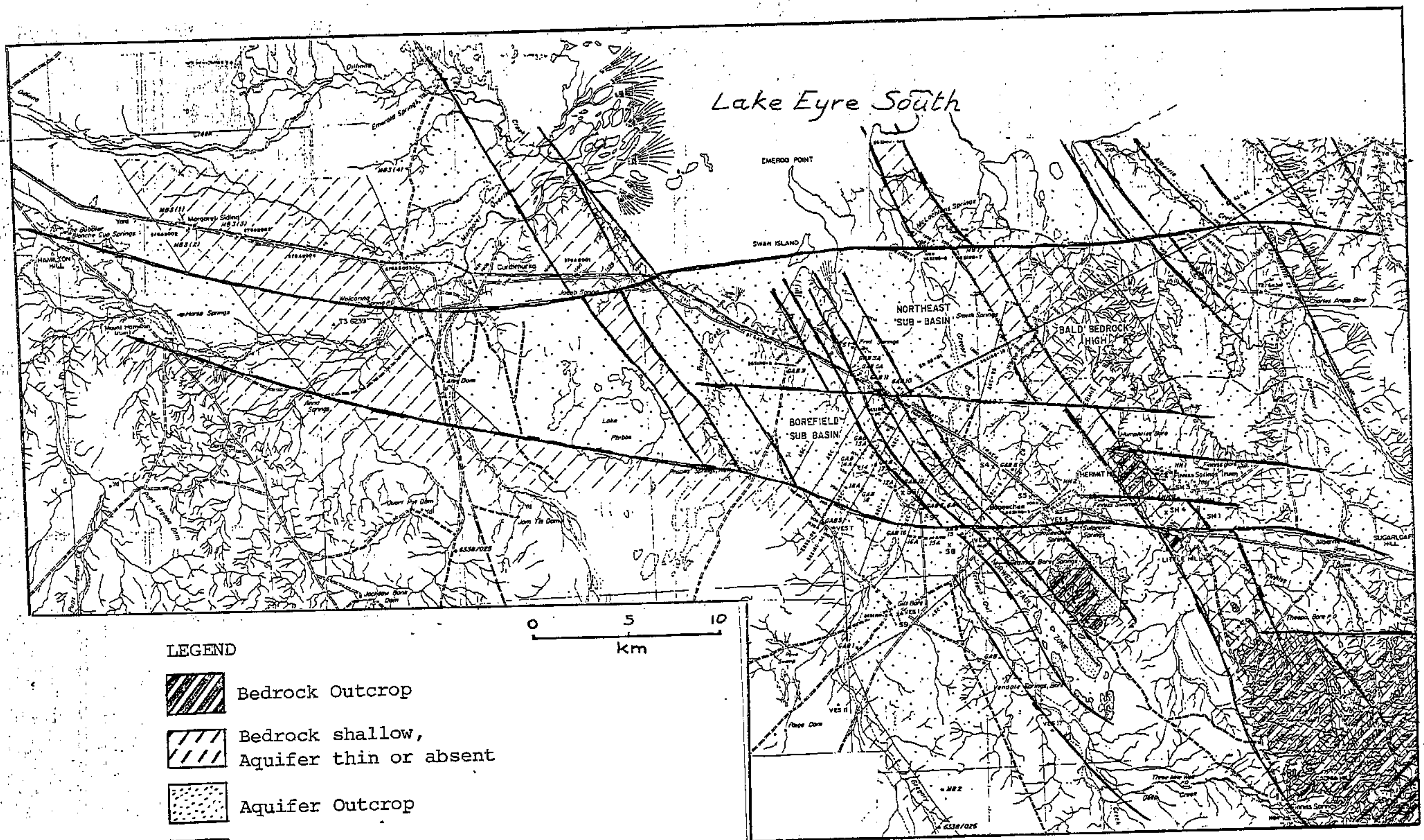
Summary Geological Sequence of Project Area

Unit	Summary Description
Quaternary Sediments	Up to 10m of dune sands, alluvium gibber plains, clay pans, mound spring and lake deposits.
Tertiary Sediments	Limestone, dolomite, sandstone, duricrust, lake sediments.
Bulldog Shale (Cretaceous)	Grey-black shale and clay, 0 to 150 m thick.
Cadna-owie Formation (Cret.), Algebuckina Sandstone (Jur.)	Sandstone aquifers of the Great Artesian Basin
Adelaidean rocks	Shales, quartzites, etc. Outcrop in Willouran Ranges

western part of the extended project area, (Ambrose & Flint, 1981, Benbow, 1983), but are not relevant to the work of this thesis. Overlying the Adelaidean basement are the late Jurassic Algebuckina Sandstone and early Cretaceous Cadna-owie Formation (Wopfner *et al.*, 1970.) The Cadna-owie consists of some silt and claystone as well as sandstone. The two units are often difficult to distinguish, and comprise the GAB aquifer in the region. They outcrop at the very margin of the basin and around the Peake and Denison Ranges, and subcrop much of the sand dune area south of Lake Cadibarrawirracanna (Ambrose & Flint, 1981). The lower, more clayey Algebuckina Sandstone is interpreted as deposited on a previously deeply weathered, stable land surface with sluggish fluviatile conditions, and the well-sorted upper part under a more vigorous fluvial regime. (Wopfner *et al.*, 1970). The Cadna-owie Formation represents the early Cretaceous marine transgression, laid down in a variety of shallow marine environments, except for the Mount Anna Sandstone Member, which is fluviatile-deltaic (ibid.). The discovery of boulders of earlier, exotic rocks in the Cadna-owie has aroused much interest over the years. Although opinions have varied, the boulders were probably carried into the area by Permian glaciation, and reworked in the early Cretaceous. They occur in the overlying, low-energy Bulldog Shale as well, and Benbow (1983) suggests a partially Cretaceous ice-rafted origin is likely, in common with several earlier workers. The area is thought to have been much closer to the south pole than at present due to continental drift.

Seismic work carried out during groundwater exploration and folding trends suggest that an ancient undulating valley and ridge topography with drainage to the north-west existed in the pre-Mesozoic surface in the project area proper. This may explain why the existence of the aquifer sands is not continuous under the full extent of shale: in addition, faulting disrupts the hydraulic continuity of the aquifer (A.G.C., 1984, also Aldam & Kuang, 1988). The extent of the J aquifer in the field area is shown in figure 1.2.9 (Cockshell, 1988).

The next unit is the Bulldog Shale (Freytag, 1966). It is a dark grey shale, fairly unconsolidated at more than a few tens of metres depth. It is carbonaceous, pyritic, and glauconitic, with some sandy beds, contains thin lenses of limestone and dolomite, particularly near



LEGEND

- Bedrock Outcrop
- Bedrock shallow, Aquifer thin or absent
- Aquifer Outcrop
- Aquifer present in subsurface
- Geological Fault

FIGURE 1.2.9 Structural Geology of the Project Area with the extent of J-Aquifer occurrence. after Cockshell, 1988.

its base, and the boulders as discussed above. The major clay mineral is montmorillonite (Lock, 1988). It contains 0.8 to 2.0% total organic carbon (McKirdy *et al.*, 1986). To the north the Bulldog Shale is overlain by the Oodnadatta Formation (Freytag, 1966), which is also predominantly a shale, but this formation is absent from the project area, presumably removed by erosion. The Bulldog Shale was exposed during the Tertiary and weathered and bleached to great depth, up to 50-60 metres as seen in the Stuart Ranges (Benbow, 1983). Several weathering events can be deduced (Jessup & Norris, 1971). In the project area this weathered zone has been removed by more recent erosion, and the shale is fairly fresh only one to a few metres below the surface. It is known to weather rapidly, and often contains secondary evaporite minerals such as gypsum, halite (sodium chloride), celestite (strontium sulphate, Williams, 1972), epsomite (magnesium sulphate), and rarer evaporites (Lock, 1988).

Tertiary sediments overlie the Bulldog Shale around Lake Eyre, but were not intersected in any of the holes drilled for this project. They are described in Wells and Callen (1986). Quaternary sediments thinly cover much of the area, varying from a few centimetres of loess to up to 10 metres of dune sand, and include the alluvium of current creek beds and flood plains, and the travertine (limestone) of some of the mound springs.

The Cretaceous sediments are cut by several sets of faults, probably reactivated from those existing in the pre-Mesozoic basement. One fault, not apparent at the surface, causes a displacement of the aquifer of 37 metres between bores only a hundred metres apart (A.G.C., 1984). The faulting divides the margin of the GAB into several sub-basins and is apparent in figure 1.2.9.

1.2.3 HYDROLOGY

1.2.3.1 Surface Hydrology - General

The Great Artesian Basin coincides for much of its extent with the surface water catchment of Lake Eyre, although the latter extends somewhat farther to the west and south. In the south-east the GAB underlies the internally draining Bulloo-Bancannia catchment and parts of the Darling River catchment (Australian Water Resources Council, 1976). There may be some contribution to the recharge of the GAB by surface waters of the Darling Basin in the aquifer outcrops in the east, but for most purposes the two are considered as separate systems, as may be said for the Bulloo-Bancannia catchment. The Lake Eyre catchment, likewise, is mostly independent of the GAB, with the same proviso regarding recharge as for the Darling (see also comments on GAB recharge that follow shortly).

The hydrology of Lake Eyre and its catchment has been of interest since white explorers first encountered the area, and has recently been researched and published in a book by Kotwicki (1986). Once thought to be dry most of the time, Kotwicki has shown that some surface water occurs on Lake Eyre on average every second year, and floods covering much of the lake every 10 or 20 years. The average annual inflow is 3.75 km^3, and has varied historically from zero for up to seven years (more commonly one, two or three) up to 39.3 km^3 in 1974. Much surface runoff is detained in river channels, lagoons, and swamps, and never reaches the terminal lake, so that actual runoff is much greater. There are some apparently permanent fresh water pools in the Cooper Creek and some other large streams in the deserts of central Australia. Some small tributaries in the tropical north-east fringe of the catchment where average annual rainfall exceeds 400 mm a^{-1} may flow every year. That part of the GAB in the extreme north (Carpentaria Basin) is overlain by some rivers that flow into the Gulf of Carpentaria each year in the summer wet season.

There are many salt lakes overlying the GAB, of which Lake Eyre is the largest. Lake Frome in South Australia is underlain by an embayment of the GAB and receives at least some water via springs (Ker, 1966, Draper & Jensen, 1976). Other salt lakes are much smaller, Lake Cadibarrawirracanna being the most notable of these in the extended project area (Fig. 1.2.2). These other small lakes are flooded by surface waters at various infrequent intervals, largely unknown due to their remoteness and sparse data.

1.2.3.2 Surface Hydrology - Project Area

The surface hydrology of the area relates to Lake Eyre South and and its catchments (Fig. 1.2.10). The major tributaries of Lake Eyre South in the project area are, from the east, Alberrie or Poole Creek, Wergowerangerilinna Creek, Screechowl and Gregory Creeks, Tinta Din-tana Creek, Priscilla Creek, and the Margaret River. They drain the Willouran and Stuart Ranges and flow briefly every few years or so (Kotwicki, 1986). In parts the creek beds are bare and salt scalded, while elsewhere they support relatively dense shrubs and small trees. They are associated with mound springs in places, such as the Wergowerangerilinna Creek and the Hermit Hill spring complex in the east, and the Margaret River and the Hamilton Hill spring complex in the west. No flow records are available. Creeks tend to be best formed on outcrop of the impervious Bulldog Shale where runoff is high during rain. There is an absence of creeks in the dunefields on Algebuckina Sandstone / Cadna-owie Formation subcrop, where infiltration of rainfall occurs more readily,

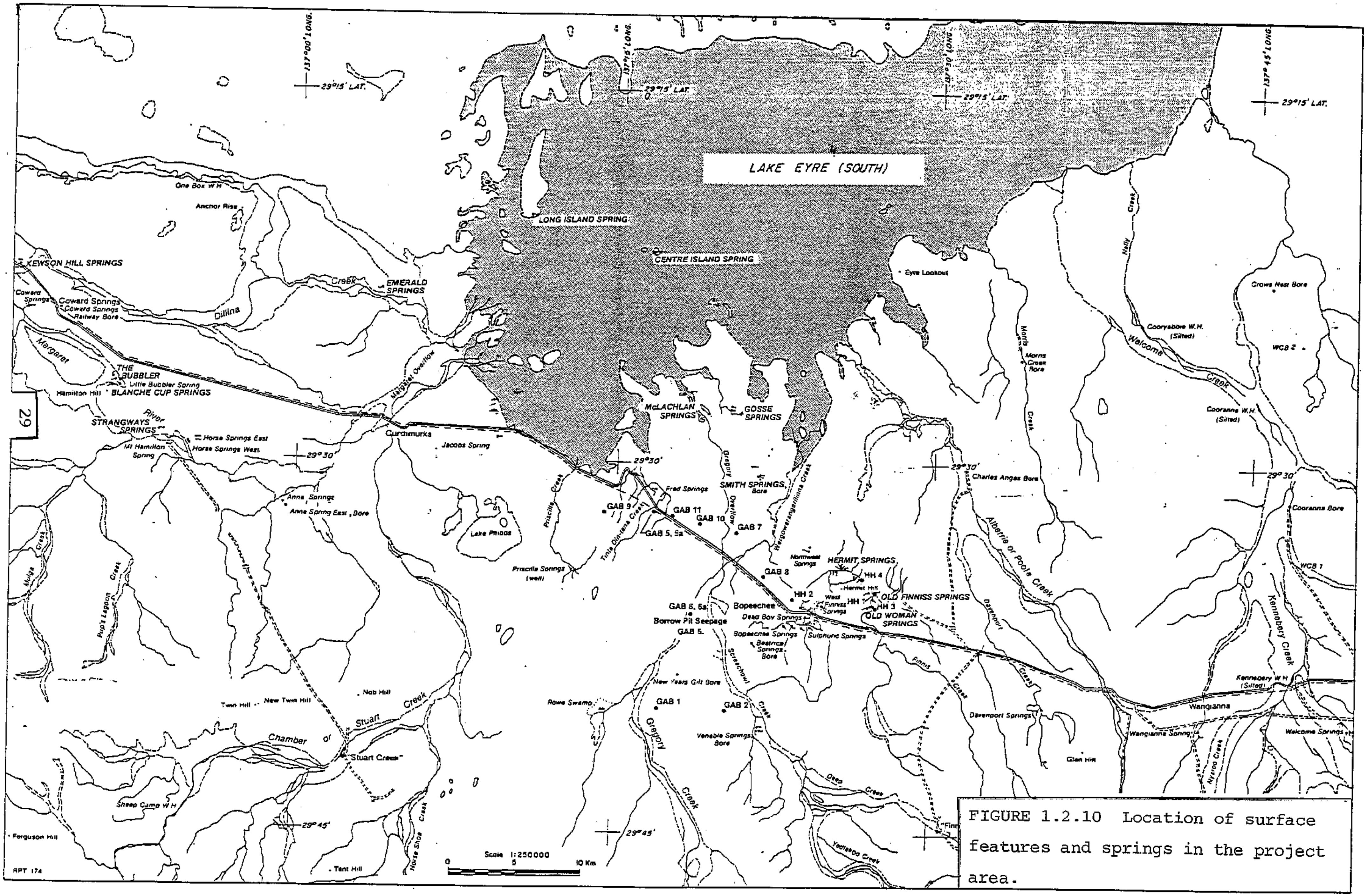


FIGURE 1.2.10 Location of surface features and springs in the project area.

possibly occasionally recharging the aquifer, although interdune claypans are common. Lake Cadibarrawirracanna in the west of the extended project area has its own roughly radial set of tributaries, the larger of which also support riparian (riverside) vegetation.

1.2.3.3 Hydrogeology - General

The best consideration of the Basin's overall hydrogeology is that of Habermehl (1980), and much of the information for this section comes from his work.

The aquifers of the GAB occur in sandstones of Triassic to Cretaceous age. They are grouped into the J and K aquifers for practical purposes (Fig 1.2.11), corresponding to those shown in cross section in figure 1.2.7. Aquifers vary in thickness from several metres to several hundred metres, and occur to depths of 3000 m (Habermehl, 1980). Recharge occurs mainly in the east, but to a lesser extent in the west (Fig. 1.2.12). Flow is dominantly to the south-west, except for west of Lake Eyre where it is to the south-east. Natural discharge is by springs and leakage to the near-surface water table from where it evaporates directly or perhaps first moves into salt lakes. The only known exit to the sea is into the Gulf of Carpentaria, for that part of the basin north of the Euroka Ridge. Artificial discharge, both deliberate and incidental, is a major component of the Basin's water balance, with natural and artificial discharge thought to be of roughly equal magnitude (ibid.).

Transmissivity values from field tests generally range from 1 to 2000 m^2d^{-1}: low values (< 10 - 20 m^2d^{-1}) predominate in the south-central and most easterly parts, higher values (10s to 100s m^2d^{-1}) in the northern and southern parts (ibid.). These values are for the most used J aquifers (Cadna-owie Formation / Hooray Sandstone and equivalents, and

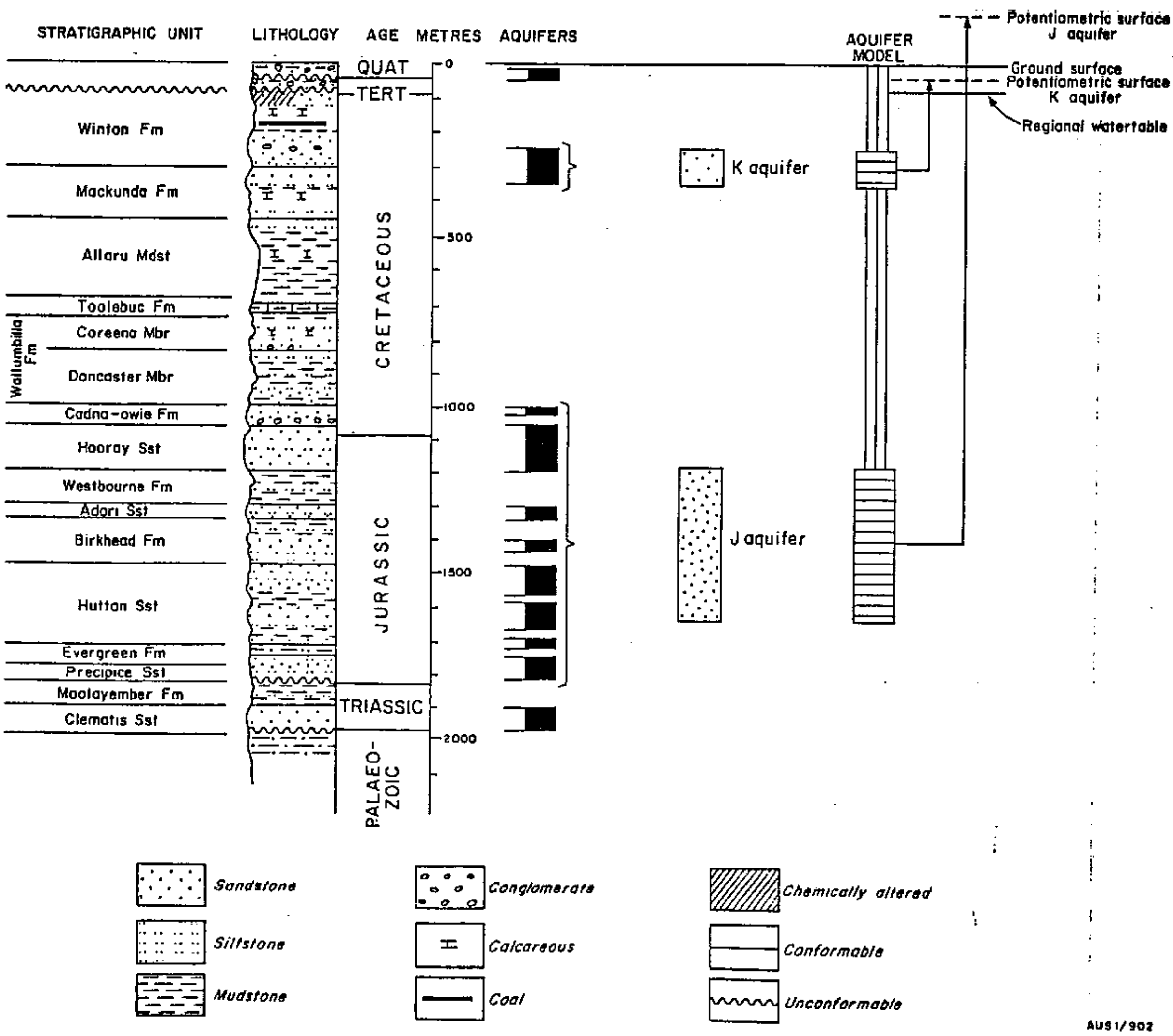


FIGURE 1.2.11 Generalised Groupings of Great Artesian Basin Aquifers (Habermehl, 1980)

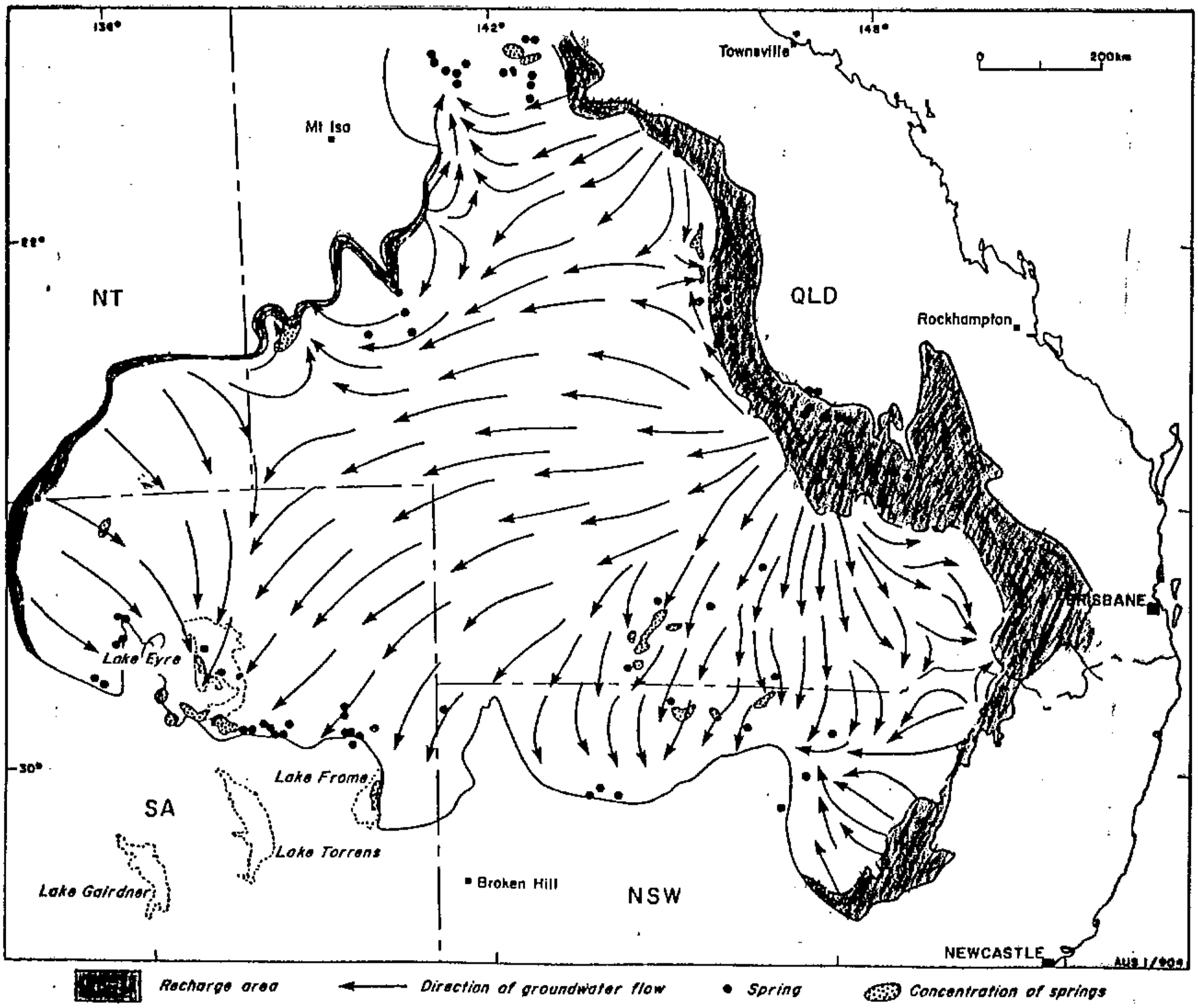


FIGURE 1.2.12 Recharge Areas, Flow Directions, and Major Spring Groups of the Great Artesian Basin (Habermehl, 1980)

the Pillinga Sandstone). Storage coefficients are around 10^{-4} to 10^{-6}, averaging about 10^{-5}, typical of confined aquifers. Vertical leakage is slight over the majority of the basin, as testified by the maintenance of artesian heads to within a few tens of kilometres of the southern margins, except where man has overused the supply in northern NSW. Some estimates in the central basin suggest vertical hydraulic conductivities (K_v) of 10^{-1} to 10^{-4} m d^{-1} (10^{-6} to 10^{-9} m s^{-1}: Audibert, 1976 (unpublished) quoted by Habermehl, 1980). Vertical leakage becomes very important near the edge of the basin where confining beds are thin, having been removed by erosion, and may be fractured. This aspect of the Basin's hydrogeology in the project area is largely the subject of this work and will be discussed at length later.

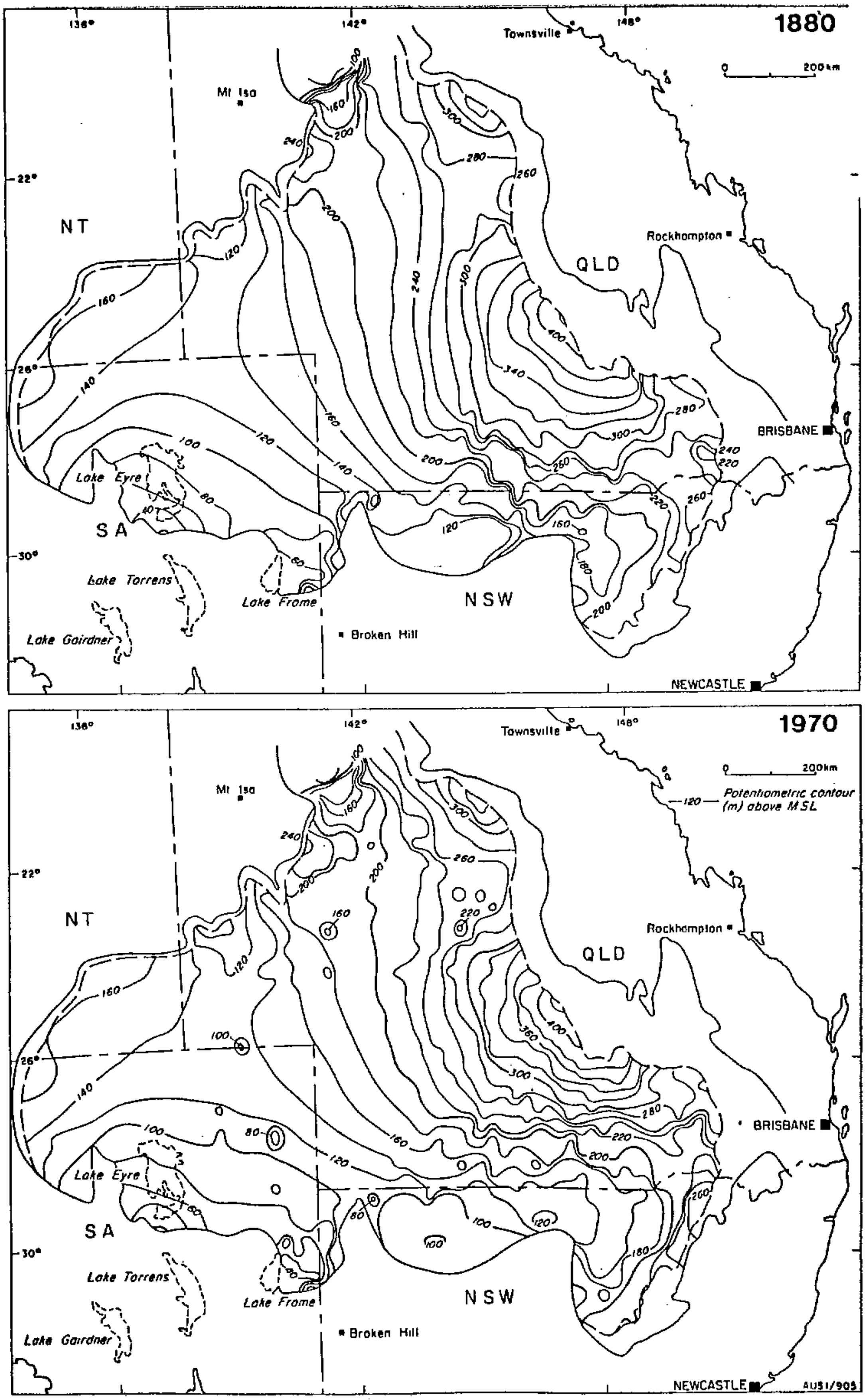
The major recharge zone of the GAB lies in Queensland. Williams and Coventry (1981) give evidence that conditions favourable for recharge occur beneath the red, yellow and grey earths that coincide with much of the inferred recharge zone, and could contribute much of the recharge required by calculated flows south-west. They quote sources mentioning the other likely water source, infiltration of surface water from creeks rising in the Great Dividing Ranges. The existence of springs and groundwater maintained pools in stream beds in the recharge zone suggest that water entry into that unconfined part of the aquifer is more than enough to feed the confined part, the recharge amount being limited by the ability of the aquifer to conduct it away (D. Armstrong, oral communication, 1988).

The head distribution of the GAB aquifers must generally be interpolated from widely spaces bores, and the exact aquifer interval(s) being measured is not always clear, especially with old bores. Pre-exploitation (1880) and 1970 potentiometric maps modified from computer modelling using the GABHYD programme (Seidel, 1980) are shown in figure 1.2.13 taken from Habermehl (1980). They reflect (in fact, were used to infer) the generally south-west flow direction. Hydraulic gradients in the J aquifer are about 1:2000, and in the K aquifer about 1:1800. Using this information together with estimates of transmissivity and porosity (0.2) allow calculation of residence time (Fig. 1.2.14, Airey *et al.*, 1983). The calculated age for water at the most distant terminus of flow near Lake Eyre is two million years.

Water temperatures are quite high in the central part of the basin, reaching 100°C due to the great depth of the aquifer. More common temperatures are 30-50°C in bores, and 20-40°C in springs. This stored heat is used to generate electricity in a trial plant at Mulka station homestead in northern South Australia.

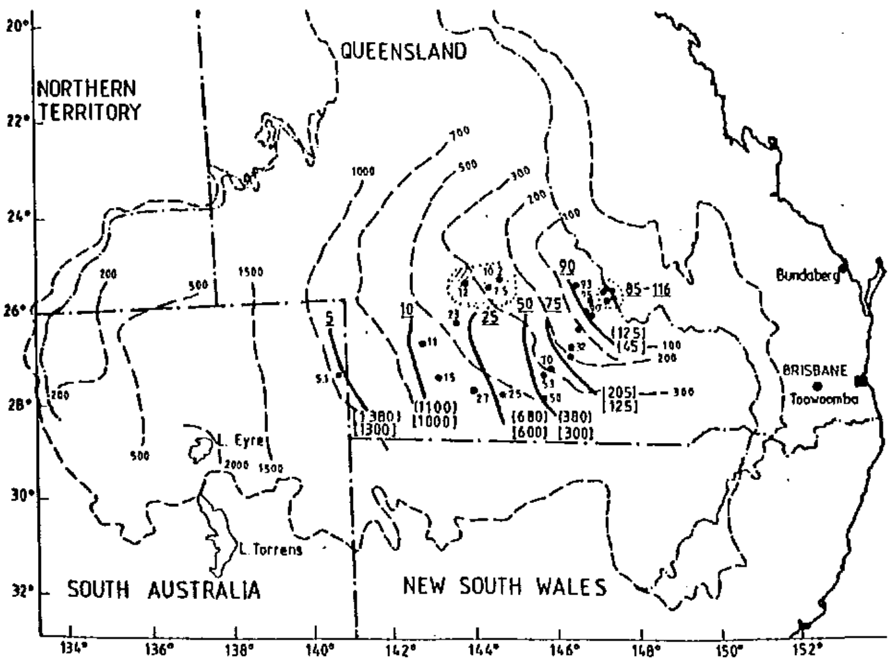
Groundwater in the most widely exploited J aquifers generally has a salinity of 500 to 1500 mg l^{-1} east of Lake Eyre. Quality tends to improve with depth of aquifer. The water chemistry is dominated by Na^{+}- HCO_3^{-}- Cl^{-} in this large eastern portion. Water in the western, often sub-artesian portion is usually more saline (1000s of mg l^{-1}) and dominated by Na^{+}- Cl^{-}- SO_4^{2-} (Habermehl, 1983), and the HCO_3^{-} / TDS ratio is a good indicator of the two types (e.g. A.G.C., 1984).

The hydrochemisty of the waters has been examined for many species by many different researchers. Habermehl (1983) considered major ions and elaborated on the east and west differences and trends. He opined that hydrochemical differences can be related to sources and direction of groundwater flow patterns consistent with hydrodynamics. Herczeg *et al.* (*in prep.*) show that the waters are in equilibrium with some clay minerals, and unsaturated with respect to calcite, amorphous silica, feldspar and smectite. The isotope hydrology has been published several times, more recent papers are Airey *et al.* (1979, 1983) and Calf and Habermehl (1984). Stable isotopes of water show a meteoric origin, with



Potentiometric maps of the main aquifers in the Lower Cretaceous-Jurassic sequence during 1880 and 1970, modified from GABHYD simulation model results, using all available potential data. Datum is mean sea level, and potentials relate to pure water at 15°C.

FIGURE 1.2.13 from Habermehl, 1980.



Superposition of the ^{36}Cl (———) and hydraulic age (-----) contours. The $^{36}Cl/Cl$ ratios (x 10^{-15}) are shown. Input values of 150 and 100 x 10^{-15} have been used to calculate ^{36}Cl isotope ages in round and square brackets respectively. An anomalous region near Adavale has been distinguished by shading.

FIGURE 1.2.14 Water Residence Times, Great Artesian Basin. Airey et al., 1983.

average $\delta D = -41.8 \pm 1.3$ and $\delta^{18}O = -6.87 \pm 0.77$ permil relative to SMOW, very close to Craig's (1961a) meteoric line $\delta D = 8\delta^{18}O + 10$ (see section 1.3, figure 1.3.1). Due to the long residence time of the aquifers, Davidson and Airey (1982) looked for a palaeoclimatic record in the δD, $\delta^{18}O$ pattern, but without success, the range being small. Airey *et al.* (1983) in their more extensive survey found an area with lighter than average isotopes that they assigned as due to recharge when global temperatures were lower up to 50 000 years ago.

Carbon isotopes have also been investigated. Carbon-14 levels above background (<35 000 years BP) only exist in small regions near eastern recharge intakes, one unexpected area in central Queensland (away from intake areas: this might represent an additional internal recharge area; Calf & Habermehl, 1984) and at the recharge zone near the Queensland/Northern Territory border (ibid.). Each of the main recharge areas have distinct carbon isotope signatures, overall $\delta^{13}C$ values range between -7 and -17 permil PDB; they can be followed downgradient where they become slightly enriched. Airey *et al.* (1983) and Calf & Habermehl (1984) assigned this enrichment to dissolution of carbonate grains of the aquifer matrix, and possibly exchange in the deeper, hotter parts. Herczeg *et al.* (submitted) however found evidence that the trend towards heavier $\delta^{13}C$ (to about -2‰ PDB), with accompanying decrease in CO_3^{2-} and HCO_3^- concentrations relative to chloride, is the result of fractionation during the reduction of CO_2 to form methane, rather than the dissolution of carbonate minerals.

The GAB was one of the first two large aquifer systems to have waters analysed for the cosmogenic species ^{36}Cl, following the development of suitable tandem accelerator mass spectrometric techniques in the late 1970s and early 1980s that enabled detection at natural abundance levels. With a half-life of 3.01×10^5 years, ^{36}Cl shows promise as a dating tool for very old groundwaters. Bentley *et al.* (1986) published results of a transect from central Queensland to Innamincka, just over the South Australian side of the border. The oldest water was calculated to be at the method's limit of 1.1×10^6 years old, and they obtained a reasonable fit to hydrodynamic age, within uncertainties (Fig. 1.2.14).

Although the waters of the GAB are of meteoric origin, the syncline does apparently act as a trap for crustal degassing. The amounts of helium-4 (Torgersen & Clarke, 1985) and argon-40 (Torgersen *et al.*, 1989) are greater than atmospheric contributions at recharge and radiogenic contributions from aquifer matrix materials alone, suggesting an additional crustal source.

The most obvious discharge points of the GAB are its springs, often referred to as mound springs due to the morphology of a great number of them. Most of the springs occur in semi-arid and arid areas and have been a focus for human activity. They have been catalogued in eleven groups (Habermehl, 1980), these groups and important individual springs are described by Habermehl (1982). They vary from slight seepages, through vegetation-clogged vents, to large springs supporting permanent creeks with flows up to 7300 m^3d^{-1}. The total discharge from all known springs (about 350) is about 0.13×10^6 m^3d^{-1} (Habermehl, 1980). The largest springs are at Dalhousie in northern South Australia (Williams & Holmes, 1978), with other major springs on the south-west margin (see below). Some springs occur where the aquifers abut impervious basement rocks, or where only thin confining beds are present. Most appear to be fault controlled; this has been confirmed by geophysical investigation (Aldam & Kuang, 1988, Cockshell, 1988). It is argued in section 1.2.5 that the greater part of the natural discharge must be by leakage.

1.2.3.4 Hydrogeology - Project Area

The extended project area covers the south-west edge of the Great Artesian Basin aquifer system, which in the west is unconfined. Western waters in the J aquifer are generally quite saline (McNally, 1980), although a ribbon of brackish water due to preferred recharge where the Bulldog Shale is thin and fractured occurs north-east of Lake Cadibarrawirracanna (Mason, 1975). Saline water also occurs in fractures or sandy lenses of the Bulldog Shale.

In the main project area artesian water of reasonable quality is available in the J aquifer, of the eastern chemical type (Na^{+}- HCO_3^{-} - Cl^{-}). Salinity increases with proximity to the basin margin and to the west, rising from 1500 mg l^{-1} TDS (total dissolved solids) to 3000 mg l^{-1} and more in just a few kilometres in the north-south direction, and tens of kilometres in the east-west (A.G.C., 1984, Fig. 1.2.15).

The hydrogeology of the project area has been studied in detail by Australian Groundwater Consultants Pty Ltd on behalf of the Olympic Dam Project of Roxby Downs. The following information comes largely from one of their reports (A.G.C., 1984), together with another by Kinhill Stearns (1984) for the same client. As described in the geology section earlier, the aquifer is absent in part and interrupted by faulting. The J aquifer cannot be distinguished clearly into the Algebuckina sandstone and Cadna-owie formation in the area. Heads vary from about 30 m to 5 m AHD (Fig. 1.2.16). Total salinity trends were mentioned above. Sulphate becomes important to the west, and Na Cl to the south and west. Bicarbonate decreases from east to west. These trends are consistent to the area's position where the east and west originating waters meet.

Many natural springs occur in the project and extended area, associated with faulting. The most important of these are the Hermit

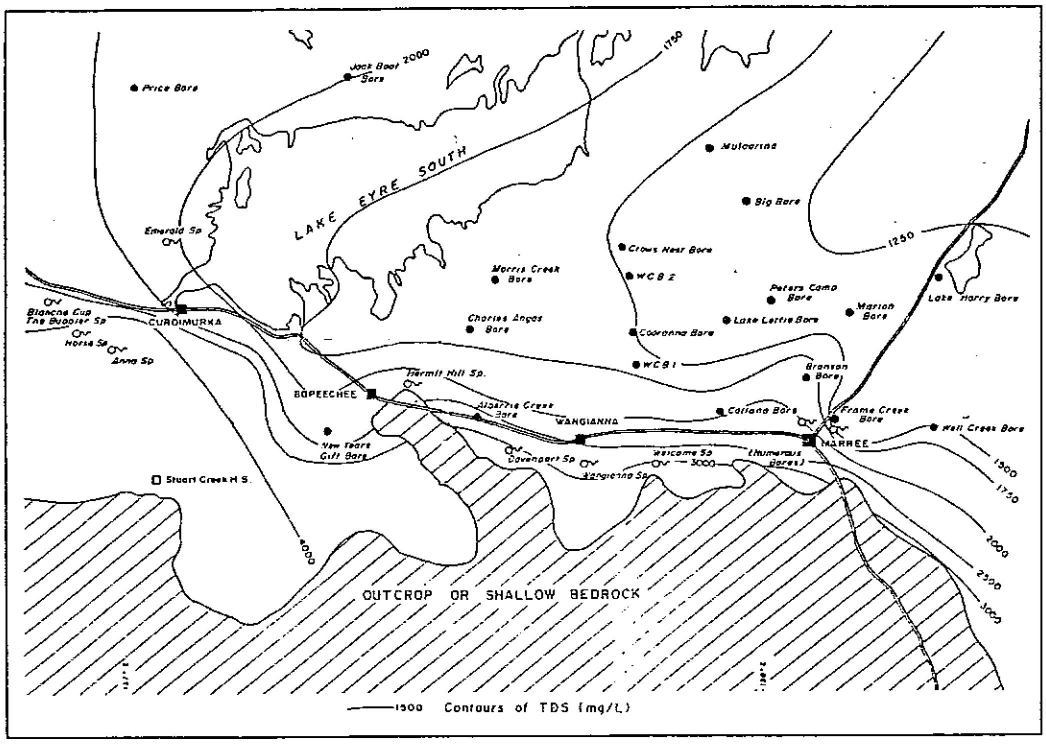


FIGURE 1.2.15 Groundwater Quality in the Project Area.
Armstrong & Rowan, 1987.

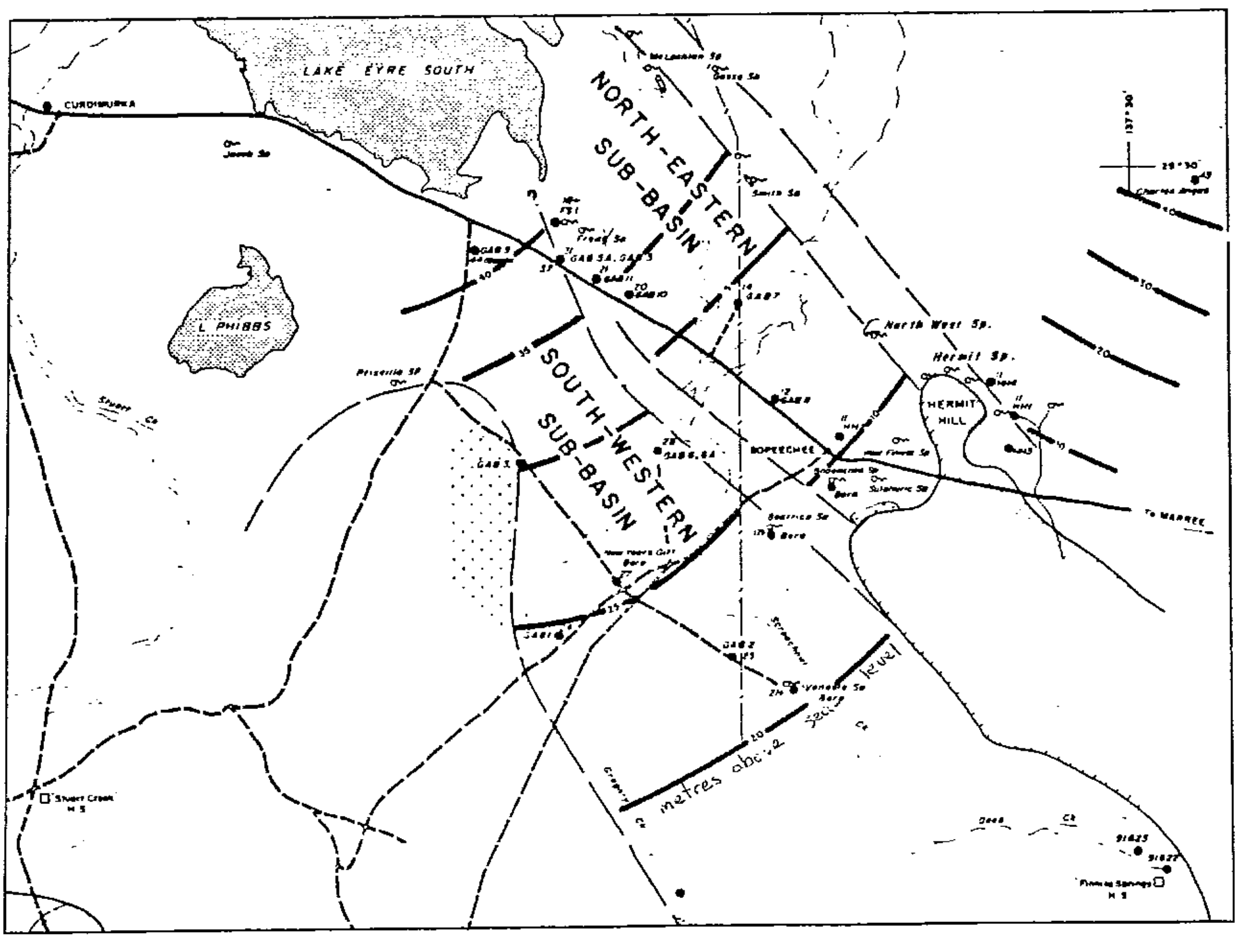


FIGURE 1.2.16 Potentiometry of the Artesian Aquifer in the Project Area, before the Roxby Water Supply Borefield Installation. A.G.C., 1984.

Hill complex in the east, and the Hamilton Hill complex in the west. The springs are described in detail by Greenslade *et al.* (1985), Kinhill Stearns (1984) and A.G.C. (1985). The flow from springs is monitored by Roxby Management Services (Olympic Dam Project), using direct methods (e.g. V-notch weirs) when possible, dye gauging (Black & Woolard, 1985) more commonly, or rarely by the amount of vegetative cover (Williams & Holmes, 1978). Flows from the springs are very variable, the individual trails from the sandy reed covered mounds such as at Hermit Hill varying in number, location and vigour over the several years of monitoring. No satisfactory explanation for all of this natural variation is yet apparent (Roxby Management Services, pers. comm., 1988). The total spring outflow between Hermit Hill and Hamilton Hill, some 70 x 20 km, is about 1000 m^3d^{-1} (A.G.C., 1984).

The aquifer is about 7 to 30 m thick where intersected at between about 50 and 300 m in the Roxby Downs borefield and nearby springs and monitoring bores. Water temperatures range from 23 to 40°C, mostly 25 to 35°C, and salinities from 1650 to 3300 mg l^{-1} TDS.

A computer model of the area and surrounds has been written and used for drawdown prediction (GABROX, A.G.C., 1984). The transmissivities used were 100 m^2d^{-1} in the south-west of the south-western sub-basin, 350 beyond, and 35 in the north-east sub-basin. Storativity was assigned to 10^{-4}, and apparent vertical permeability to 2.7 x 10^{-4} m d^{-1}, an average fairly high for such a fine-grained aquitard, to allow for fractured and silty parts of the shale and give the most reasonable behaviour of the model compared to measured data. A revised model was produced in 1989, but the work has not been obtained by the author (A.G.C., 1989).

A water table exists in the Bulldog shale at a depth of one to several metres. Where the shale is tight, the water table is just that surface where the pressure of pore water is equal to atmospheric, but where the upper part of the shale is fractured, a water table aquifer may be said to exist. Where encountered, this upper water is extremely saline, with salinities of greater than seawater (32 000 mg l^{-1}). Its chemistry is dominated by sodium chloride, and salinities reach 200 000 mg l^{-1} (A.G.C., 1984, and this work).

1.2.4. UTILIZATION

1.2.4.1 General

The groundwater of the GAB is a major resource over a vast tract of Australia (Habermehl & Seidel, 1979). Its springs were used by aboriginal peoples and local fauna, then by graziers for domestic and stock supplies. Their distribution in South Australia dictated the route of the first north-south railway and telegraph. Some of these uses continue today, and the more accessible and spectacular springs have become tourist attractions.

The first shallow flowing bore was drilled near Burke, NSW, in 1878, and many followed near springs on the basin margins, then inward. The first deep flowing bores were drilled in response to drought in central Queensland in 1887 (ibid.). Not many bores extend to the J aquifer in the centre of the basin because of the great depth, supplies being pumped from the shallower K aquifer. The generally Na^{+}- HCO_3^{-} rich groundwater, combined with regional soils in the arid zone is considered unsuitable for general irrigation despite its freshness (Shepherd, 1978). Western waters however are saltier (stock quality only) and corrosive (ibid.). About 3100 flowing and 20 000 subartesian bores currently tap the waters. Artesian bores are up to 2000 m deep, but average about 500m, and may exceed 10 000 m^3 d^{-1} yield, though the average is much less. Non-flowing bores are generally equipped with windmills and yield about 10 m^3 d^{-1} (Habermehl, 1980). The overall withdrawal in the 1970s was about 1.5 × 10^6 m^3 d^{-1}, down from 2.0 × 10^6 m^3 d^{-1} in 1918, when water was released from elastic storage (ibid.). The lessening is due to drastic head falls of up to 80m in the South-east, with a subsequent fall in flows. Heads and flows appear to have levelled off and reached a new equilibrium between recharge and discharge at 3.0 × 10^6 m^3 d^{-1}. That water not withdrawn by bores is consumed by natural discharge at springs and by diffuse leakage.

The use of artesian water from the GAB has been characterized by very low efficiency. Flowing bores were and still may be let run at maximum flow, and water distributed by open channels for tens of kilometres. It is estimated that nearly 90% of this water is wasted by seepage, transpiration and evaporation (Habermehl, 1980). Towns would seem highly efficient by comparison. Some uncontrolled (perhaps abandoned) bores sustain large pools and swamps in the desert. Since 1977 the South Australian Department of Mines and Energy has brought under control over 90 bores, and pastoral users are encouraged to install tanks and pipelines (SADME, 1984b). It is hoped to extend this practice to the other involved states, with government assistance/obligation to land users (X. Siebenhaler, oral comm., 1988).

Various small towns such as Birdsville (Qld) and Marree (S.A.) make use of GAB water. Two major users have recently appeared. The older is the petroleum/natural gas extraction industry in the north-east of South Australia (Moomba) and adjacent fields in Queensland and New South Wales. The other is the Roxby Downs mining venture in South Australia, which since the late 1980s draws its supplies from the southern margin of the GAB near Lake Eyre. This last use will be further discussed in the following section.

1.2.4.2 Project Area

As elsewhere in the GAB, early use of mound springs was by fauna and aboriginal peoples. The first pastoralists (late 1800s) used springs as watering points, fencing some off and conducting water to troughs (e.g. Fred Springs), and the north-south rail link followed the line of springs as the only water supply in the desert. Some early bores were drilled adjacent to or within springs (e.g. Venables), and then at other points (e.g. New Years Gift), and fencing usually left to fall into disrepair. As well as for pastoral and homestead use, bores were drilled for a town water supply at Marree, and to supply the various sidings on the now abandoned narrow gauge line railway.

The largest user is the Olympic Dam Project and its township of Roxby Downs. The Olympic Dam copper - uranium - gold mine is located some 110 km south of its borefield. The large underground ore deposit (Roberts & Hudson, 1983, 1984; Youles, 1984) was discovered in 1975, and commenced production in 1988/89 (Newton *et al.*, 1988). When it became apparent that water could not be obtained locally, nor economically piped in from the south, groundwater exploration was centred around New Years Gift bore south of Bopeechee on the margin of the Great Artesian Basin (Armstrong & Rowan, 1987).

The prospect of a large (up to 33 $Ml\ d^{-1}$ in the long term, 15 in the first few years) extraction from the margin of the GAB raised some concerns that the biologically significant mound springs might be dried up (e.g. McKinnon, 1987). A special licence was issued by the South Australian government, the amount of water allowed to be taken defined by the borefield induced drawdown at the edge of a designated area (SADME, 1984a, Armstrong & Rowan, 1987). The area appears to have supported the early requirement of 9 $Ml\ d^{-1}$ ($9000\ m^3d^{-1}$) without undue hazards

(Waterhouse & Armstrong, 1990). The Hermit Hill spring complex is protected by the low permeability fault zone separating the south-west (borefield) and north-east sub-basins (Armstrong & Rowan, 1987).

Water is piped to Olympic Dam and used untreated for processing, supplemented by local saline groundwater from the mine shaft and a local borefield, with use of recycling (Nadebaum & Amiconi, 1986). The potable requirements of the mine and township are provided by an electrodialysis desalination plant (Rymill, 1985). The town has been designed with water efficiency in mind (Zwar, 1988).

1.2.5 IMPORTANCE OF DIFFUSE DISCHARGE

As implied by the GAB water balance of Habermehl (1980), the assumed loss of water by leakage (diffuse discharge) at 1.4×10^6 m^3 d^{-1} is about 10 times that by natural point discharges (springs) at 0.13×10^6 m^3 d^{-1}, and similar to the total artificial point discharge (bores) of 1.5×10^6 m^3 d^{-1}. While springs and bores are quite obvious, leakage is not. While there is probably some downward leakage into underlying geological basins, at the first approximation this is probably matched by leakage back into the GAB, contributing some salt to the system (D. Armstrong, oral comm., 1988).

Leakage that causes discharge is upwards, ultimately to the water table aquifers. From there it moves directly or laterally into salt lakes and evaporates (Habermehl, 1980) (or, just possibly, in the south east, into the Darling River system, Jolly, 1989), or evaporates directly from the water table where that is shallow. This leakage is likely to be greatest from the edge of the Basin, where confining beds are thin. Because of the large area involved, only low leakage rates are required: over the entire basin, the rate would be

$1.4 \times 10^6\ m^3\ d^{-1} / 1.7 \times 10^{12}\ m^2 = 8 \times 10^{-7}\ m\ d^{-1}\ (3 \times 10^{-4}\ m\ yr^{-1})$.
A leakage of 1 mm yr^{-1} would account for all the lost water in an area of $0.5 \times 10^6\ km^2$, or 30% of the basin, or a leakage of 5 mm yr^{-1} over only 6%. Such small rates of leakage as these can be removed by evaporation with the water table a couple of metres below the surface. This argument will be pursued in chapter 6.

The Roxby Downs water supply draws water from a terminus of GAB flow where discharge of water, largely by leakage, is occurring. In one sense, it is harvesting this discharge by intercepting water before it can leak away and evaporate, plus taking some water drawn in by increased hydraulic gradients from further out in the basin. Diffuse discharge then gives an estimate of the sustainable yield of the utilized sub basin. If the amount of water withdrawn reverses the upward leakage through the shale, downward leakage of poor quality water would adversely affect the water supply (Armstrong & Rowan, 1987). Current preferred policy by the users is therefore to let the bores flow naturally, or restrict pumping so that the water level does not fall much below ground level, and so maintain the area's upward hydraulic gradient.

Due to the low hydraulic conductivity of the shale (except where it is strongly fractured) wherein the water table exists in the project area, local leakage is most likely consumed by local evaporation. In such an arid climate shallow water-table evaporation is often amenable to measurement by use of stable isotope and chloride profiles (Barnes & Allison, 1988), and hence the generation of this project. A direct measure, or at least confirmation of the order of upward leakage, can be compared to the quite indirect measure of the same by hydraulic computer modelling.

1.3 STABLE ISOTOPE HYDROLOGY

The stable isotopes of hydrogen and oxygen, in water, are of wide use in many aspects of hydrology, as evidenced by various book chapters (Freeze & Cherry, 1979, Drever, 1972) and proceedings of symposia (e.g. Perry & Montgomery, 1982, and many I.A.E.A. (International Atomic Energy Agency) publications). Measurement of these isotopes at high resolution became possible following the development of dual-inlet, multiple faraday collector mass spectrometers in the late 1940s and the 1950s. The approximate relative abundances of stable isotopes in sea water are (Montgomery & Perry, 1982)

^{1}H	99.98%	^{16}O	99.76%
^{2}H (or D)	0.016%	^{17}O	0.016%
		^{18}O	0.20% .

Of the oxygen isotopes, only ^{16}O and ^{18}O are of hydrologic interest. Hydrogen-2 is commonly called deuterium, symbol D. In the earth sciences, the amount of heavy isotope in a sample is usually expressed in delta notation, as a permil (‰) difference from a standard,

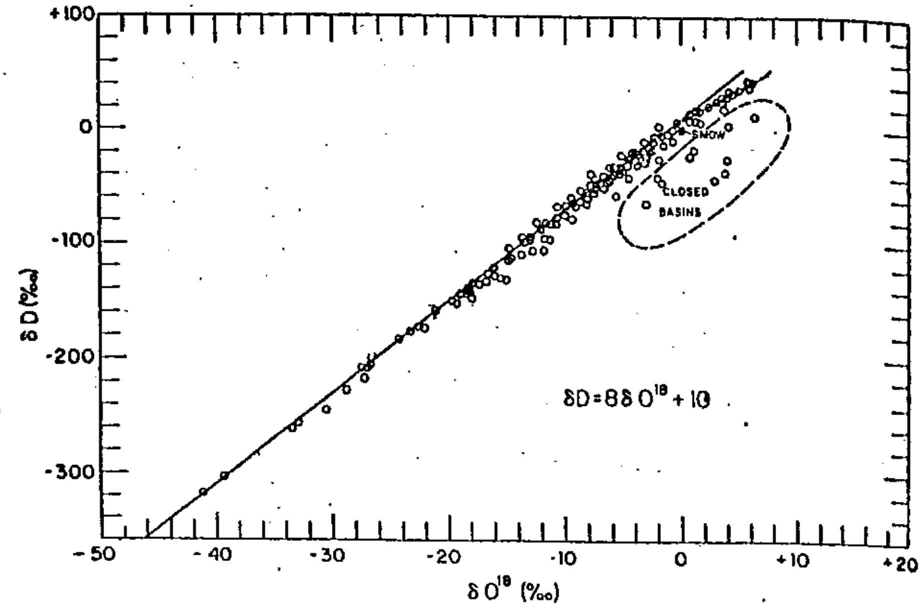
$$\delta i = [(R_{unknown} / R_{standard}) - 1)] \times 1000$$

where i is the isotope of interest, and R the heavy-to-light isotope molar ratio (D/H or $^{18}O/^{16}O$ in this case, which correspond to $HD^{16}O$ / $H_2{}^{16}O$ and $H_2{}^{18}O$ / $H_2{}^{16}O$ in natural waters). The accepted international standard is V-SMOW (Vienna- Standard Mean Ocean Water, kept by the I.A.E.A. at Vienna, often referred to simply as SMOW), which has absolute deuterium and oxygen-18 contents of 155.7 and 2005.2 ppm (Gonfiantini, 1984). The delta values are, by definition, 0.00, and the world's oceanic waters are close to this.

Water containing heavy isotopes is fractionated during phase changes and chemical reactions, i.e. the isotope ratios are changed. Theory dealing with the behaviour of isotopes is presented in section 2.1. The isotopic signature of natural waters therefore often contains information about such changes or reactions, from which inferences concerning hydrologic processes can sometimes be made. Largely because of the lower vapour pressure of molecules containing the heavy isotope, one very important fractionation is that occurring between liquid water and its vapour during evaporation and condensation. At equilibrium, water vapour is lighter by about 70 and 9‰ (δD and $\delta^{18}O$ respectively) than liquid water, with the actual amount (fractionation) varying slowly with temperature (Friedman & O'Niel, 1977). The ratio of about 8 between the equilibrium fractionation of δD and $\delta^{18}O$ is a significant one. If the isotopic composition of natural precipitation over much of the globe is plotted with δD and $\delta^{18}O$ on the y and x axes (Craig, 1961a, Fig. 1.3.1), the slope of the resulting line is 8. Typical ratios also exist for some non-equilibrium processes. Evaporation of water through a dry layer of soil (mulch) causes a ratio of fractionations of as low as 2, compared with evaporation from open water bodies where the ratio is about 5 (Allison, 1982).

The relationship between the stable isotope composition of hydrated minerals and water in equilibrium with them is often different, for example with gypsum the ratio between hydrogen and oxygen fractionations is about $-\overline{5}$ (Gonfiantini & Fontes, 1963; Fontes & Gonfiantini, 1967; Sofer, 1978). Some exchange reactions between water minerals containing one of the elements in water but not the other (mostly oxygen) changes the content of that isotope alone.

Because deuterium and oxygen-18 may be subject to different processes and fractionations, additional information is often obtained by examining both δD and $\delta^{18}O$ in a set of water samples. However, where the



Deuterium and oxygen-18 variations in rivers, lakes, rain, and snow, expressed as per millage enrichments relative to "standard mean ocean water" (SMOW). Points which fit the dashed line at upper end of the curve are rivers and lakes from East Africa.

FIGURE 1.3.1 Craig's Meteoric Line of World Rainfall Isotopic Composition. Craig, 1961a.

relationship between the two is constant over a whole set of samples, little or no additional information is gained by the analysis of more than one isotope.

1.4 MEASUREMENT OF EVAPORATION FROM A WATER TABLE

1.4.1 INTRODUCTION

Evaporation from the soil, and to a lesser extent from shallow water tables has been recognized as an important component of the water balance for many years, particularly in an agricultural context. Understanding evaporation (and evapotranspiration) from a water table found application particularly in irrigated agriculture, where shallow saline water tables are common. Excessive water-table evaporation leads to salinisation of soils, because of the salts left behind, lowering their usefulness. The term critical water-table depth has been used to describe the level below which, in given circumstances, the water table should be maintained in order for agriculture to be sustained (Talsma, 1963; Szabolcs *et al.*, 1972; Oswal & Kaker, 1987). Talsma (1963) gives curves relating maximum evaporation to depth to water table for several soil types (Fig. 1.4.1, from Gutteridge *et al.*, 1970). In general to maintain agriculture the critical water-table depth is about one to two metres, being deeper in clayey soils; this corresponds to evaporation rates of about one mm d^{-1}.

In natural settings, evaporation from water tables can be an important sink for water in enclosed basins, particularly in arid zones. Evaporation of groundwater from salt lakes has been estimated by various methods. Where the overall water balance of a basin may be calculated,

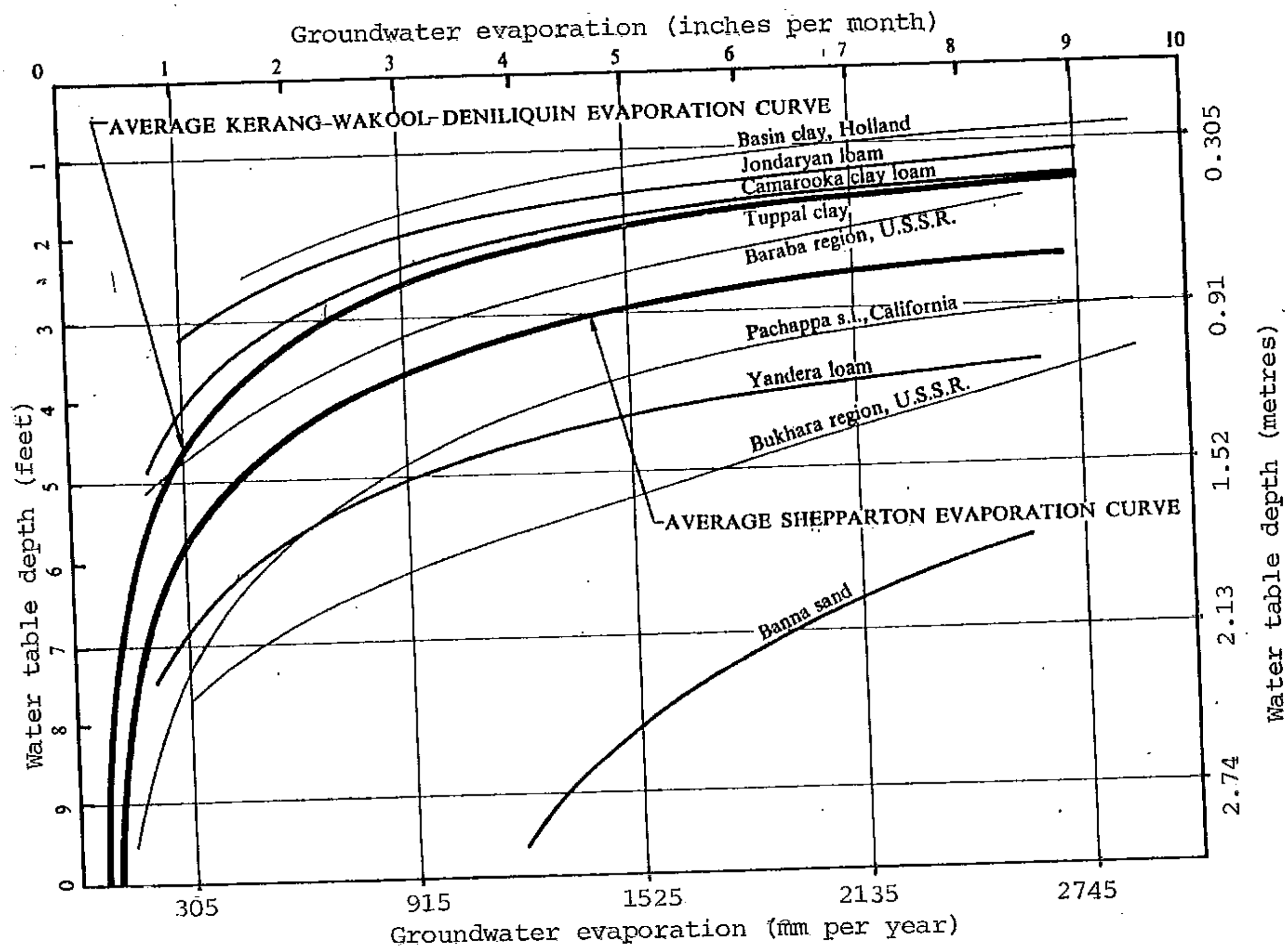


FIGURE 1.4.1 Evaporation versus groundwater depth curves of Talsma (1963)

the component due to water-table evaporation can be calculated by difference (e.g. Schmid, 1985, Lake Torrens; Woods, 1983, Lake Bumbunga; Jacobson & Jankowski, 1989, Lake Amadeus: all in Australia). Various physical methods will be outlined in the following subsections. Following laboratory work (Zimmermann *et al.*, 1967; Allison *et al.*, 1983), the vertical profiles of stable isotopes in water (deuterium and oxygen-18) were applied to estimate the evaporation from shallow water tables in dry Lakes Frome (Allison & Barnes, 1983, 1985) and Torrens (Schmid, 1985) in Australia, and depressions in the eastern Sahara (Christmann & Sonntag, 1987), shallow water tables near Lake Eyre (this work), and deeper water tables in Algeria (Fontes *et al.*, 1986). Profiles of other solutes can sometimes be used, e.g. Ullman (1985), using chloride and bromide, and Allison and Barnes (1985), using chloride.

All the solute profile work just mentioned is restricted to unvegetated areas. The presence of plants violates one or more of assumptions involved in such calculations, and other means have proved difficult. Wronski (1986) obtained some uncertain measurements in sandy terrain near Perth, Western Australia, and some progress is being made addressing the issue in the Murray Basin (Allison *et al.*, in prep.; Brunel *et al.*, 1989; see section 1.4.2.5 for some details). The consumption of groundwater by vegetation has importance in the combating of dryland salinity (Greenwood *et al.*, 1985; Jolly, 1988).

1.4.2 ANALYTICAL AND PHYSICAL METHODS

1.4.2.1 Analytical

Applying Darcy's law to describe the movement of water in soils (convection), and using analytical descriptions of the relationship between unsaturated hydraulic conductivity K_{unsat}, and matric suction, ψ, Gardner (1958) provided several steady-state analytical solutions relating evaporation rate to depth to water table. These were applied to laboratory experiments with columns of soil (Gardner & Firman, 1958). Further, more generalised solutions have been published by Warrick, (1988). Talsma's curves (fig 1.4.1), based on such solutions, may be used to obtain an order-of-magnitude value for water-table evaporation in an irrigated, vegetated context. They do not compare well with evaporation rates obtained by tracer methods in natural desert conditions, particularly on salt lakes where salt crusts may by present, reducing evaporation by reduction of the vapour pressure gradient between the evaporation front and atmosphere, and high albedo resulting in less adsorption of solar heat (sections 1.4.4, 3.3).

Many attempts have been made to quantify non-steady-state evaporation as it proceeds in a soil column (e.g. Black *et al.*, 1969; Walker *et al.*, 1988). Most of the water evaporated comes from that stored in the soil profile rather than that transmitted up from a water table. Such non-steady-state approaches would be useful to describe a situation such as when a water table, initially at the surface, progressively becomes deeper as evaporation proceeds.

1.4.2.2 Hydraulic Measurement and Calculation

Calculation of vertical water flux (driven by evaporation) in soil is possible if suction may be measured as a function of depth and the relationships between K_{unsat}, ψ, and θ (water content) are known (Gardner, 1958). The measurement of suction *in situ* is difficult, particularly at high suctions, and is subject to considerable field variability (e.g. Greminger *et al.*, 1985). Use is made of tensiometers up to about 85 kPa suction (Marshall & Holmes, 1988), or filter papers from about 1 to 100 kPa (Greacen *et al.*, 1987, 1989). The nature of the $K_{unsat} - \psi$ relationship is likewise difficult to measure, varying over orders of magnitude of both quantities, and is very difficult to quantify accurately. As a consequence, measurement of water-table evaporation by direct hydraulic measurement and calculation is impractical, at least in the field area of this project. A suction profile can at least serve to indicate whether water movement in a soil is up (evaporation) or down (infiltration; e.g. Marshall & Holmes, 1988, pp79-80, reworking data of Greminger *et al.*, 1985).

The vapour flux may also be calculated from the relative humidity gradient in the dry, upper part of a soil profile. Relative humidity can be calculated from matric and osmotic suctions if they are known, and the flux calculated using the air-filled porosity and an effective gaseous diffusivity (equivalent to hydraulic conductivity - section 2.4).

1.4.2.3 Lysimeters

Lysimeters are buried pots or drums containing soil intended to match that around them, for which the water balance may be determined, usually by weighing. They may be as small as flower pots and pulled up and weighed entire (Chen & Bowler, 1988), to several cubic metres in volume, weighed on an in situ underground balance (Marshall & Holmes, 1988). They tend to be expensive but can give quite accurate results of infiltration if well prepared.

Small lysimeters are used by Chen and Bowler (1988) to try and estimate evaporation from the generally dry playa Lake Amadeus in Central Australia. The pots are set in the generally dry bed of the lake, and periodically pulled out and weighed. The pots are sealed at the base however, and so not in hydraulic connection with the shallow water table at about 0.3 - 0.4m, so it is difficult to see that the evaporation measured is that from the water table; however, only an abstract of that work has been examined. Lysimeters are not considered by this author as likely to be useful to give a representative estimate of groundwater evaporation in natural semi-arid to arid environments where total fluxes are small, though in more humid or irrigated areas they could be of good use.

1.4.2.4 Micrometeorological measurement

Various micrometeorological methods are available to measure evapotranspiration, and occasionally applied where evaporation from the water table may be important. Pike (1970) used such techniques (heat balance and Bowen ratio) over a coastal playa (sabkhah) in Arabia, and estimated an annual groundwater evaporation of about 0.7m, extrapolated from just two day-long measurements several months apart. In a more intensive study that includes a review of such methods, they were used in Libyan playas (Menenti, 1984), and also by Schmid (1985) at Lake Torrens in South Australia. Eddy correlation, another micrometeorological technique, has been used on Algerian playas (J.-P. Brunel, pers. comm.).

Due to the inherent inaccuracies, it is not practical to measure very low evaporation rates (less than some tens of millimetres per year) by such methods.

Airborne measurement of total evaporation by instrumented light aircraft is possible, using micrometeorological principles (Hacker & Schwedtfeger, 1988). This has been applied as one of several techniques in south-western New South Wales in an area where groundwater evaporation was considered significant (Allison *et al.*, in prep.), but to the current level of development provides more qualitative than quantitative information, and cannot distinguish water source (be that stored soil water or groundwater).

1.4.2.5 Measurement of Transpirational Loss from Groundwater

Many plants (phreatophytes) are known to draw at least some of their water requirements from groundwater. While various techniques exist that can estimate the transpiration of vegetation, it is very difficult to distinguish that component coming from groundwater from that coming from stored soil moisture.

A ventilated chamber method (Greenwood & Beresford, 1979) has been used to measure evapotranspiration from vegetation where groundwater is thought to supply some of the water (Farrington *et al.*, 1989; Allison *et al.*, in prep.). The contrast in evapotranspiration between sites with the water table at different depths can give an idea of the groundwater component.

Potential exists to use stable isotopes of water, where a contrast in isotopic composition exists between soil and ground waters, to differentiate the components by looking at the isotopic composition of sap water (White *et al.*, 1984; Richardson, 1987; Brunel *et al.*, 1989), although further work is required to provide confident estimates of actual transpiration of groundwater (Brunel *et al.*, 1989).

Using the solute balance of the capillary fringe and daily fluctuations of the water table under *Banksia* woodland and pine plantations north of Perth, Western Australia, Wronski (1986) obtained very rough estimates of groundwater transpiration of 100 mm a^{-1} by *Pinus pinaster* and much less by *Banksia* woodland (water table at 2 to 3 m).

1.4.2.6 <u>Remote Sensing</u>

Both temperature and albedo may be determined by aircraft and satellite mounted instrumentation, and so an estimate of evaporation obtained by heat balance, with ground calibration (van de Griend & Gurney, 1988). With regards to evaporation of groundwater, this has been attempted by Menenti (1984) in Libya, and Schmid (1985) in South Australia. The former used Landsat, HCMM and Meteosat data over 36,000 km^2 of Libyan desert that included many playas, and suggested an average evaporation of 220 mm yr^{-1} over 3800 km^3 (including 194 km^2 of playa lake surface), given without an estimate of confidence level or error. The figure implies an excess of evaporative losses over recharge for the aquifer system at the ratio 1.42 : 1, and a nett depletion of 40 mm yr^{-1} drawdown over the entire system which appears to have been losing water since the last pluvial period 11,000 years ago. A rate of 220 mm yr^{-1} over such a large area, which includes the surrounds of playas as well as the well defined playa lakes (which only occupy 5.1% of the stated evaporating area) seems excessive compared to later studies in the Sahara and arid Australia. These show on playas, by various other methods, rates of 50 mm yr^{-1} (Lake Torrens, Schmid, 1985, by water balance; he used satellite data for extension of point measurements rather than raw estimates), 170 mm yr^{-1} (Lake Frome, Allison & Barnes, 1985), 4 to 700 mm yr^{-1} in East Saharan depressions (Christmann & Sonntag, 1987), 9 to 27 mm yr^{-1} (Lake Eyre, Ullman, 1985), 49 mm yr^{-1} (Lake Amadeus, Jacobson & Jankowski, 1989), and 60-100 mm yr^{-1} (Lake Bumbunga, Woods, 1983), and low rates on the surrounds, of 1 to 2 mm yr^{-1} in the West Sahara (Zouari *et al.*, 1985, Fontes *et al.*, 1986) and about 0.5 to 4.5 mm yr^{-1} near Lake Eyre by this work. Nevertheless, remote sensing is a promising approach for extending point measurements made on the ground, and will no doubt be applied increasingly in the future. The interpretation of satellite data to obtain quantitative information is not straightforward, as evidenced by the calculations of Menenti (1984), and as the sophistication of both satellite sensing and interpretation improve, reliable areal estimations will be forthcoming.

1.4.3 TRACER METHODS

1.4.3.1 Stable Isotopes of Water

The use of stable isotopes of hydrogen and oxygen in tracing the movement of water in the unsaturated zone has been reviewed by Allison *et al.*, 1984, and more recently by Barnes and Allison (1988). With regard to the estimation of evaporation, Zimmermann *et al.* (1967) showed that in a saturated sand undergoing evaporation, the enrichment of deuterium in the soil water decreased exponentially with depth, with the decay length approximately proportional to the evaporation rate. This pattern is explained by evaporative enrichment of the heavy stable isotopes and simple convection-diffusion theory. The work was extended to unsaturated soils by Münnich *et al.* (1980) and Barnes and Allison (1983), and the characteristic lowering of slope of the deuterium - oxygen-18 relationship noted (Allison, 1982). Later non-isothermal effects (Barnes & Allison, 1984) and non-steady state were considered (Walker *et al.*, 1988). The theory to the extent relevant to this study is developed in section 2.2. Although work on profiles of stable isotopes has been underway for two decades or more, and quantitative estimates of discharge (and in different situations recharge) can sometimes be made, the understanding is incomplete, particularly the estimation of effective diffusivity for a given soil and water content (Chapter 2, particularly section 2.1.4). As put by Barnes and Allison (1988, p172),

> Without some effort devoted to understanding the actual variation of the diffusivities for different water contents and materials, the stable isotope technique will remain a rather blunt tool, failing to achieve the precision potentially available, not only for estimating evaporation rates, but also for determining soil physical mechanisms operating during evaporation, for example.

1.4.3.2 <u>Chloride</u>

As a conservative, commonly occurring solute in natural waters, chloride has proved a very useful and widely used tracer in hydrologic studies, particularly for recharge studies (Allison, 1988; Allison & Hughes, 1983; Eriksonn & Khunakasem, 1969). While the overall increase in dissolved solids in a groundwater discharge area compared to inflowing water may be used to obtain a rough idea of evaporative discharge (e.g. Passmore & Woods, 1986), chloride profiles in the unsaturated zone may be used for direct point estimates in the manner of stable isotopes if the appropriate boundary conditions are met. Allison and Barnes (1985) obtained evaporation estimates from chloride profiles beneath a thin salt crust at Lake Frome, and Ullman (1985) from beneath a thicker salt crust at Lake Eyre. The presence of the mineral Halite (NaCl) ensures that a constant concentration boundary condition is met, enabling convection-diffusion steady-state theory to be used (section 2.2.2). While such conditions are met in such salt pans, elsewhere the lack of this constant concentration boundary condition has limited the utility of a chloride profile for obtaining hydrologic information (Fontes *et al.*, 1986). In many of the sites drilled as part of this project, the mineral halite was present near the top of the profile so that evaporation estimates could be made (see Chapter 5). In addition to chloride, Ullman (1985) also used the vertical distribution of bromide in his single hole in the halite crust of Lake Eyre to estimate evaporation.

1.4.3.3 Previous Studies

The contributions of early workers to the development of a convection-diffusion theory for the estimation of evaporation from isotope profiles (adapted for chloride) will be credited in the section on theory (2.2); the published applications will now be briefly examined.

Lake Frome, South Australia. Allison & Barnes, 1983, 1985

Lake Frome is one of the large dry lakes in the semi-arid to arid north of South Australia. Mean annual rainfall is about 175 mm. Five cores were taken by push-tube, along a transect from the western shore onto the lake proper. Water and chloride content were determined, and water extracted from the muddy sediments by azeotropic distillation with toluene (c.f. section 3.2.2). The result of a single hole using deuterium only were published in 1983, with full results (including oxygen-18 and chloride) in 1985. Estimates of evaporation were obtained from the shape of the isotope profile beneath the evaporating front, the depth of the evaporation front, and the shape of the chloride profiles, the first and last with estimates of error, based on the fit of the actual profiles to theory. The first estimates were preferred, giving values of 90, 90, 120, 150, and 230 mm yr^{-1}, with errors of about 20%, going from the shore towards the centre of the lake, with water table varying from about 0.7 to 0.0 m. Variations in water content on the effective diffusivity of the isotopes were accounted for, but a constant tortuosity (formation factor; sections 2.1.4, 2.2) of 0.66 (Penman, 1940) used. By current thinking this may be too high for such clays (acknowledged by Barnes & Allison, 1988), so that the estimates are too high, perhaps by as much as a factor of two; this would be in accord with the South Australian Department of Mines and Energy's recent opinion that the initially published numbers are too high to be in accord with the water balance of the closed basin (D. Armstrong, pers. comm.). The work represents however a breakthrough in arid zone water resources evaluation, by allowing direct measurement of a component of the water balance unmeasurable by previous techniques.

Southern Tunisia. Zouari, 1983; Zouari *et al.*, 1985

This study took three cores by dry rotary coring to depths of about 20 m, just above the water table, in a locally low area overlying a major confined aquifer. Average annual rainfall was about 200 mm. Two holes out of three proved amenable to the Barnes and Allison (1983) approach, yielding an evaporation estimate of 1 mm yr^{-1} from oxygen-18 profiles. They were drilled in sands and silts, with small quantities of gypsum, except for one gypsum layer in hole 3. Water was extracted by vacuum distillation at 55 to 70°C, and a tortuosity of 0.66 used in calculations.

Southern Algeria. Yousfi, 1984; Yousfi *et al.*, 1985; Fontes *et al.*, 1986

This work was carried out near the town of Béni Abbès in the Great Erg sand desert of the Sahara, where average annual rainfall is only about 40 mm, either side of a valley containing the Wadi Saoura. Three holes were drilled, and two analysed for stable isotopes. One of these showed a readily interpretable stable isotope profile, yielding an evaporation estimate of 2 mm yr^{-1}, with a water table at about 10m. The other showed a profile indicating evaporation, but disturbed near the surface, so that an evaporation estimate could not be made. The disturbance was considered to be due to an infiltration event in the recent past, such that the profile had not regained steady state. The Penman tortuosity of 0.66 was used.

Lake Torrens, South Australia. Schmid, 1985

As part of a doctoral study on this playa lake, four soil profiles were taken and analysed for deuterium, and convection-diffusion theory applied to estimate evaporation. The profiles showed a broad evaporation front, and using Penman's tortuosity of 0.66, evaporation was estimated to be 30 mm yr^{-1}, similar to the value of 50 mm yr^{-1} obtained by water balance methods.

Lake Eyre, South Australia. Ullman, 1985

From a single core through the salt crust of Lake Eyre, Ullman used the vertical distributions of chloride beneath the crust, and of bromide within the halite crust to estimate evaporation. The latter version of the method relied on the exclusion of Br^- from the crystal structure of halite to set up the upper boundary condition. The estimate of evaporation from bromide was 27 mm yr^{-1}, that from chloride 9 mm yr^{-1}, the difference being assigned to the characteristic times of the profiles, about 0.2 and 20 years respectively, i.e. the bromide gave a seasonal (summer) estimate, and the chloride a long term value that included a major flooding of the lake some nine years before the core was taken. He implicitly used a tortuosity of 1.0, considering effective diffusivity to be the product of the porosity and free solution diffusivity alone, "In the porosity range of these sediments" (0.48), following work done in marine sediments (Ullman & Aller, 1982). In the light of other terrestrial work, this would seem to mean the evaporation measurements were over-estimated by a factor of two or three.

East Saharan Depressions. Christmann, 1986 (thesis, not obtained for review); Christmann & Sonntag, 1987

Some fourteen cores were taken in the Western Desert of Egypt, from which seven estimates of evaporation were made from isotope profiles, varying from 700 to 4 mm yr^{-1} with groundwater depths of 0.35 to 2.2 m. Penman's tortuosity was again used. As with other North African work, evaporation rates were applied as aquifer discharge to explain the slow drawdown of the regional aquifers over the past few millennia, the system not yet having come to equilibrium with the very low recharge rates in effect since the region became arid a few thousand years ago.

Southern Algeria. Colin Kaczala, 1986; Allison *et al.*, 1987.

Continuing the French work at Béni-Abbès, this work involved five holes, analysed for isotope profiles of water liquid and vapour, and carbon dioxide. Three were interpreted using convection-diffusion theory to estimate evaporation, although at least one appears to be far from in steady-state (G.R. Walker, pers. comm.). Evaporation estimates varied from 1 to 7 mm yr^{-1} from water tables deeper than 7 m, using Penman's tortuosity. The isotopic relationship between water vapour and liquid within the profiles showed mostly equilibrium, with the exception that the position of the peak values of heavy isotope in the vapour samples did not correspond to those in liquid water in two holes, being above by about 0.8 m in one, and below by about 0.5 m in another (data incomplete however). Oxygen in CO_2 was in equilibrium with that in both vapour and liquid water below the zone where vapour transport was important (a little below the water isotope peak), but usually not so above. Laboratory experiments demonstrated that even at high soil suctions, the equilibrium fractionation between soil water and water vapour appeared to be the same as for a free water surface.