

INTERACTION BETWEEN GROUNDWATER AND SURFACE SYSTEMS IN NORTHERN VICTORIA



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INTERACTION BETWEEN GROUNDWATER AND SURFACE SYSTEMS IN NORTHERN VICTORIA

By Dr Phillip G. Macumber

DEPARTMENT OF CONSERVATION AND ENVIRONMENT

FOREWORD

This publication by Dr Phil Macumber is the culmination of 25 years' work on groundwater / surface water interactions in Victoria, written around the occurrence, character and movement of water and salt in the landscape. It is a landmark publication providing a technical basis for understanding the processes and trends occurring within the groundwater systems. While initially written as a PhD thesis, it has been subsequently expanded to incorporate more specific details on technical aspects of salinity control.

The work examines the hydrological and environmental instability in the landscape and details the hydrodynamic, hydrochemical and geomorphological interrelations which are both responsible for determining the nature and distribution of salt in the landscape and are in turn themselves modified by that interaction. The report is particularly significant in that it focuses jointly on groundwater and surface waters. Historically, these two systems have been studied independently by researchers from different disciplines.

The clear message to emerge is that the management of streams, wetlands, irrigation areas, and environments in general, requires an understanding of the role of groundwater systems and their influence on surface processes. It follows, that for the development of meaningful management strategies, surface water systems and groundwater systems must be considered as being part of a single integrated water resource. With the advantage of hindsight, there is no doubt that some of today's salinity problems in Victoria could have been avoided if this close association had been appreciated when irrigation areas were being initially developed.

The study also provides a model for the formation and distribution of copper, lead, zinc and other sulphide ore bodies in salt lake settings. A better understanding of the formation of these ore bodies has far reaching implications for the exploration of new deposits.

While the area examined is northern Victoria, the principles and processes apply generally for much of Australia. It is a contribution which even before its publication is widely acknowledged as providing a significant advance in the understanding of the movements of salt and water in the landscape.

Dr Macumber has made an enormous scholarly contribution to our knowledge of an important element of the Australian continent. In very recent years, this has had a major impact upon the response of Australian society to salinization. It is a model of scientific discovery and its diffusion, which stands as a splendid example to all. The Department of Conservation and Environment is proud to have made this work available to a wide audience.

PREFACE

Across much of Australia we are today experiencing a wide-ranging and fundamental shift in environmental equilibrium brought about by the impact of European man on what was previously a largely untouched landscape.

A major factor in this shift has been the large-scale clearing of trees from both the highland areas and the plains. This has led to vastly increased runoff from the catchments, resulting in accelerated erosion and sedimentation in differing parts of many drainage basins. However, perhaps the most drastic changes stem from increased infiltration, which has mobilized large volumes of salt stored in groundwater catchments and raised water tables over extensive areas, causing severe salinization in many parts of southern and eastern Australia.

Late last century, salinity first appeared as a significant problem in the Kerang irrigation district, situated in the Loddon Valley of northern Victoria; some eighty years later it is now accepted that a major environmental problem exists not only in the Loddon Valley but in virtually all the irrigated lands and much of the dryland country of northern Victoria. Today, virtually all of the lower central Loddon Valley has a high water table, and is varyingly saline (Macumber, 1978a); in the adjacent Shepparton irrigation region water tables are steadily rising, and are within two metres of the surface over an area of 100,000 hectares, i.e., about one fifth of the region (Trehalla and Webster, 1978). Further west, in the Mallee, rises in saline water tables are evidenced by the salinization of depressions (Rowan, 1971; Macumber, 1978c, 1980, 1983a) and stock dams, and by the cutting of roads. Dryland salting is also widespread on the slope country (Jenkin, 1979), and rapid deterioration is occurring along the edge of the Riverine Plain, adjacent to the highlands.

While the recent appearance of salt in the landscape has had a marked effect on the environment, saline zones were already in existence in parts of Victoria prior to European settlement. For instance, salt was once regularly harvested in the Kerang district, while in the Mallee there are large numbers of salt lakes and large areas of saline gypsum flats. Excellent examples are seen in the Pink Lakes, Lake Tyrrell, and Kow and Raak plains areas. An examination of these areas shows that highly saline zones are an essential component of the northern Victorian landscape, and that the extent of the naturally saline area has varied considerably throughout the ages.

The widely observed effects of the shift in environmental equilibrium since the coming of the Europeans provides perhaps one of the most complete records of how the interplay of the various physical and biological factors influences the landscape. Since these processes are still strongly active, it permits a rare opportunity to study the various hydrological thresholds influencing this equilibrium. This is not simply an academic exercise: much effort has already been spent on treating the symptoms of change without any firm evidence that the salinity problem is being contained, or even slowed, let alone reversed. A clear understanding of the basic processes involved is essential for an overall view of what is happening and why.

While studies of the presently occurring events are essential, the broader temporal perspective provided by the study of palaeoenvironments is also of great value. Based on the stratigraphic, biostratigraphic, geochemical, and geomorphic evidence of earlier naturally occurring cycles of environmental change, palaeoenvironmental studies greatly broaden our perspective of present-day changes, and may, in certain instances, provide added credence not only to our interpretations of presently occurring events but also to our predictions of the direction and extent of future environmental changes.

With a view towards gaining a fuller understanding of the influence of regional groundwater systems on the environment, this work embarks on a broad study of aspects of the hydrogeology of northern Victoria, with emphasis placed on the groundwater systems and their interactions with surface waters.

ACKNOWLEDGEMENTS

I wish, firstly, to thank Dr D. Spencer-Jones, former Deputy Secretary of the Victorian Department of Minerals and Energy, for sanctioning my pursuit of this project at the Geology Department of the School of Earth Sciences, Melbourne University; and for his encouragement throughout my involvement with Quaternary studies in general, and salinity investigations in particular, during my employment with the Department of Minerals and Energy.

I am greatly indebted to my supervisor, Dr R. J. McLaughlin, and to my colleague, Dr C. R. Lawrence, who gave invaluable advice, valued support, and continued encouragement throughout the project. I owe a deep debt to my wife, Jenny, who was, as always, patient and understanding, and who provided great assistance throughout the work. I thank Dr John Paterson, Director-General of the Department of Water Resources, for his support for the publication of this work.

Advice and data were obtained from many sources, including both governmental organizations and private bodies. I wish to acknowledge these valuable contributions, and thank their personnel. They include the State Rivers and Water Supply Commission - Messrs G. Gordon, G. Earl, R. H. Thorne, D. Ife, and B. Foley; Messrs K. Garland, G. O. Jones, and J. Newell of the Victorian Department of Agriculture; Dr J. J. Jenkin, and Messrs J. N. Rowan and P. Dyson of the Victorian Soil Conservation Authority; Messrs W. H. Williamson and D. Wooley of The Water Resources Commission of New South Wales; Messrs J. B. Firman, S. R. Barnett, P. A. Rogers, and R. Callen of the South Australian Department of Mines and Energy. ICI (Aust) Ltd, through Mr M. Melvin, provided invaluable information on the results of their drilling program in the Sea Lake area, while other bore hole data were supplied by Sedimentary Uranium N. L., through Mr K. D. Donaldson. I am especially grateful for the assistance and information supplied by Messrs T. Riley, G. Walker, and B. Britton of Cheetham Salt Pty Ltd, and for their hospitality.

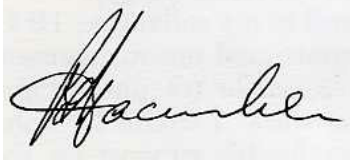
This thesis represents the culmination of a lengthy period of research, during which there were many long and stimulating discussions with colleagues in the Department of Minerals and Energy. I wish to thank Bill Esplan, former Head of the Groundwater Section; John Talent, now Associate Professor of Geology at Macquarie University; Richard Lakey, John Leonard, Mick Malone, Young Nahm, Mark Reid, Mirko Riha, Andrew Shugg, Graham Stewart, and Rob Williamson from the Groundwater Section; John Knight, former Director of Geological Survey; and Peter Kenley, former Assistant Director. Rob Williamson and Richard Lakey also advised on the computer programs used in analysing chemical data. I am greatly indebted to Tony Paskevicius, Trevor Ingram, and Doug Stone, without whose assistance the program could not have been carried out.

I thank my colleagues in the Salinity Unit of the Department of Water Resources-Margaret Kelso, Rae Moran, Campbell Fitzpatrick, and Peter Sutherland for their valuable advice on conceptual aspects of the salinity problem included in the updated version of this work, and Rob Skinner for his support.

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sequences and from the fluviatile-lacustrine sequences at Lake Wirrengren. During my travels in the Mallee I was shown generous hospitality by Gloria and Harold Macarthur, Bert Curtis, Duncan Thompson and Tim and Brian O'Sullivan and families.

I wish to thank the various groups within the Department of Minerals and Energy who have supplied facilities and support during the project: Mr R. Mealin of the State Laboratories supervised the chemical analyses, Mr C. DeFina and Mr L. Coughlan of the Drilling Branch supervised the drilling associated with the project, Mr W. E. Bennet and Miss J. Daley of the Draughting Branch supervised the draughting, and Mr R. Fox carried out the surveying. Finally, I wish to thank Noel Siver for her meticulous work in editing and proof reading the thesis, and Monika Evers for getting it into a publishable format.

A handwritten signature in black ink, appearing to read 'Macumber', is written on a light-colored background.

Phil Macumber

1. INTRODUCTION

1.1 Nature and Scope of the Study

This study, dealing with groundwater systems in northern Victoria and their interactions with surface systems, grew out of an early involvement with the salinity problem in the Loddon Plain of northern Victoria. At that time, salting was generally assumed to be a post-irrigational problem, related essentially to the physical characteristics of the surficial sediments and the soils. However, the presence of a large number of nearby salt lakes, coupled with evidence of artesian conditions within the deeper aquifers, pointed to the possibility that while salinity was directly related to irrigation practice, there was a fundamental hydrological condition present that was also influencing the environmental response to irrigation.

Before any meaningful hypothesis could be proposed which linked the salinity problem, the salt lakes, and the underlying artesian aquifers, it was first necessary by means of a drilling program to establish the structural and stratigraphic framework for the drainage basin, and thereby determine the nature of the regional groundwater flow systems. The drilling program provided not only an understanding of the hydrogeology of the Loddon Basin (Macumber, 1978a) but also gave an insight into the Tertiary evolution of the Murray River drainage basin, of which the Loddon River is a tributary (Macumber, 1978b). Furthermore, since the Loddon Plain spans the boundary between the marine and fluvial parts of the Murray Basin, the program also provided data on the stratigraphic relationships between the marine and nonmarine sequences. While the relationship between the regional groundwater flow systems and the distribution of groundwater discharge zones was being established, a major hydrological event occurred in the 1973-75 period which revealed the delicately balanced hydrological equilibrium which existed between the groundwater systems and the surface water systems. This event was the sudden response of the deep piezometer network to sheet flooding of the Loddon Plain. At much the same time, it became clear that pressure levels had been continuously rising in the corresponding deeply buried regional aquifer in the neighbouring Campaspe Valley since before 1900, at a uniform rate. This permitted the prediction of the eventual establishment of a regional discharge zone in the Campaspe Valley which would greatly exacerbate the salinity problems already existing in the area. The 1973-75 rise in pressure levels in the Loddon Valley led to a 20 km upbasin migration of the hinge line between the regional groundwater recharge zone and the discharge zone. At Bears Lagoon, in the newly expanded discharge zone, a detailed study of the consequent discharge phenomena was undertaken.

Following the completion of the original study, new data from the Goulburn and Murray valley systems became available showing rising trends in the regional aquifers similar to those observed elsewhere on the plains (Macumber, 1984, 1985, 1986). Since this more recent work contains critical supporting evidence for hydrologic change on the Riverine Plain, data from it has been included in this report. In addition, a better conceptual understanding of the applicability of salinity control options was brought about by the division of the State into 15 salinity provinces and the development of the salinity management option trees - (SMOTS) (Macumber and Fitzpatrick, 1987). In order to tie the salinity work together, a subsection on salinity has been included in Chapter 4.

The second part of this study is devoted to the characteristics of groundwater flow in the Mallee region, in order to compare regional groundwater flow systems in non-marine sequences with those in the adjacent unconfined Parilla Sand aquifer of marine origin. In the aeolian Mallee landscape of northwestern Victoria, surface drainage is virtually absent, and downbasin flow is essentially via the groundwater systems, the uppermost of which is developed in the unconfined late Tertiary Parilla Sand aquifer. On passing downbasin, the water table gradually approaches the surface to become artesian in the northern Mallee, where ground-water outcrop is evidenced by the many saline lakes and gypsum flats scattered throughout the linear dunefields, and by the very large groundwater discharge complexes, the boinkas (Macumber 1980).

Salinity is also a major problem in parts of the northern Mallee where many previously grassed and treed depressions have suddenly gone saline, reflecting a gradual rise in the saline regional water table which has developed in the Parilla Sand. In this work, emphasis was again placed on the discharge areas; the chemistry of the salt lakes is shown to be determined by that of the regional groundwaters. Some unusual features of the groundwater flow system are highlighted, especially with respect to the hydrochemistry of the aquifers and the springs; for instance, the widespread occurrence of low pH groundwaters, the similarity of the regional groundwater with sea water, and the high salinities reached by groundwater brines.

The last part of this study deals with aspects of the hydraulics of the regional groundwaters and Tyrrell Basin groundwater brines; it provides an insight into brine formation in a continental setting, and the subsequent influence of the groundwater brines on both regional groundwater flow in the vicinity of salt lakes, and on lake-groundwater interactions.

1.2 Location and Physiography

The study area covers the western part of the Riverine Plain in Victoria extending westward across the aeolian landscape of the Mallee region of northwestern Victoria - an overall area of about 50,000 km² (Figure 1.1). Within the Riverine Plain, the greater part of the study was carried out on the Loddon River drainage basin, with some additional work on the Campaspe River drainage basin farther to the east. The Campaspe and Loddon rivers are the most westerly rivers which flow into the Murray River from the south. They rise in the southwestern extremity of the Great Dividing Range, which there forms a relatively low physiographic divide (less than 100 m above sea level) between the northflowing streams of the Murray Basin and the southflowing streams draining towards the Otway Basin.

Westwards of the Loddon River plain lies the Mallee region with its largely aeolian landscape. Streams which pass northward across the Mallee do not reach the Murray River but, instead, flow into terminal lake systems within the Mallee. The streams discussed in this report include the Avoca River which normally terminates at Lake Bael Bael and the Marshes;

The Avoca River distributaries, Lalbert Creek and Tyrrell Creek, which flow into Lake Timboram and Lake Tyrrell respectively; and the Wimmera River, which normally terminates at Lake Albacutya, but which occasionally continues northwards through the Wyperfeld lake system towards Wirrengren Plain (the final lake on the Wimmera River system, last filled in 1923). For the Mallee region, greatest attention has been paid to the regional groundwater discharge areas of the northern Mallee, including principally the Sunset Country of far northwestern Victoria, and the Tyrrell Basin. (It had previously been proposed that Lake Tyrrell should be used as an evaporating basin for saline water which could be pumped from irrigation districts on the Riverine Plain).

1.3 Climate

Climatic data was obtained from the Commonwealth Bureau of Meteorology and was collected by a number of recording stations scattered throughout the study area. Apart from the more southerly highland areas, the region is essentially semiarid; there is a tendency towards hot and dry summer months when the area comes under the influence of dry variable winds, which are associated with the high pressure (anticyclonic) belt. The rainfall period is in winter and is controlled by winter westerlies and the jet stream (Gentili, 1971).

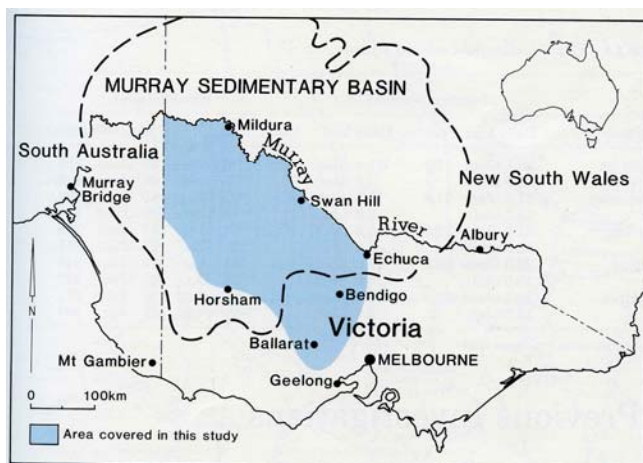


Figure 1.1 – Study area – northwestern Victoria

1.3.1 Rainfall

Rainfall is closely related to topography in the study area. The heaviest rainfall, which averages about 1060 mm per annum, occurs on the catchment divide, at the headwaters of the Loddon River, but decreases westward along the divide to about 700 mm in the catchment area of the Avoca River. On the Riverine Plain and in the Mallee annual rainfall is generally less than 500 mm, falling to about 250 mm in the far northwest of the Mallee. Northwards, across the Loddon Plain, rain varies from about 450 mm nearer the highlands to 370 mm near Kerang (Table 1.1).

Most of the rainfall occurs in winter, especially in the catchment area and on the Loddon River divide: the wettest month (June) has up to three times the rainfall of the driest (February). Within the Mallee seasonal contrast in the rainfall is not sharp, as shown by Ouyen and Sea Lake data: each having an annual rainfall of about 340 mm, a driest month of 20 mm (January), and a wettest month of 36 mm and 35 mm respectively (October).

1.3.2 Temperature

Like rainfall, temperature is closely related to physiography. The average maximum temperature is lowest (about 17°C to 18°C) in the southeastern highland region, but it rises to about 31°C to 32°C on the plains and in the Mallee (Table 1.1).

TABLE 1.1 - Climatic data from selected stations

Station	Mean temperature (°C)			Rainfall (mm)			
	Daily Max.	Year	Daily Min.	Year	Max.	Min.	Annual
Ballarat	25.5 (Jan) 9.9 (Jul)	17.3	11.4 (Jan) 3.2 (Jul)	7.3	*74 (Aug) **77 (Aug)	38 (Jan) 32 (Jan)	719 719
Rochester	31.2 (Jan) 13.1 (Jul)	21.8	15.8 (Feb) 3.2 (Jul)	8.9	*47 (May) **47 (Jan)	20 (Jan) 30 (Dec)	460 470
Kerang	31.3 (Jan) 13.8 (Jul)	22.5	15.5 (Feb) 4.0 (Jul)	9.5	*34 (Aug) **38 (Oct)	12 (Jan) 21 (Jan)	365 367
Ouyen	32.5 (Jan) 15.0 (Jul)	23.5	15.6 (Jan) 4.2 (Jul)	9.6	*31 (Aug) **36 (Oct)	12 (Jan) 20 (Jan)	346 337
Mildura	32.1 (Jan) 15.3 (Jul)	23.6	16.9 (Jan) 4.3 (Jul)	10.3	*25 (Aug) **29 (Aug)	11 (Feb) 18 (Apr)	271 294

* = Median value, ** = Mean value

1.4 Previous Investigations

A number of reports on the geology and hydrogeology of northwestern Victoria have been written since systematic drilling for groundwater commenced late last century. A summary of this early work was included in a report by Gloe (1947), who comprehensively reviewed the bore hole data from the Mallee and Wimmera regions. An outline of the physiography of northwestern Victoria was provided by Hills (1939). Johns and Lawrence (1964) first discussed the geology and groundwater resources of the southeastern Riverine Plain in Victoria, while Lawrence (1966) gave a comprehensive account of the stratigraphy and structure of the Mallee. The relationship between salinization and geomorphic processes was first investigated by Macumber (1968), who later (Macumber, 1969) showed that the Parilla Sand extends eastward into the Loddon Plain and is marine. Lawrence (1975) carried out an integrated study of the hydrogeology of the southern Murray Basin.

During the course of the investigation, certain aspects of the work were published, especially those which were most relevant to the role of groundwater systems in salinization. Data from these papers are referred to throughout the text.

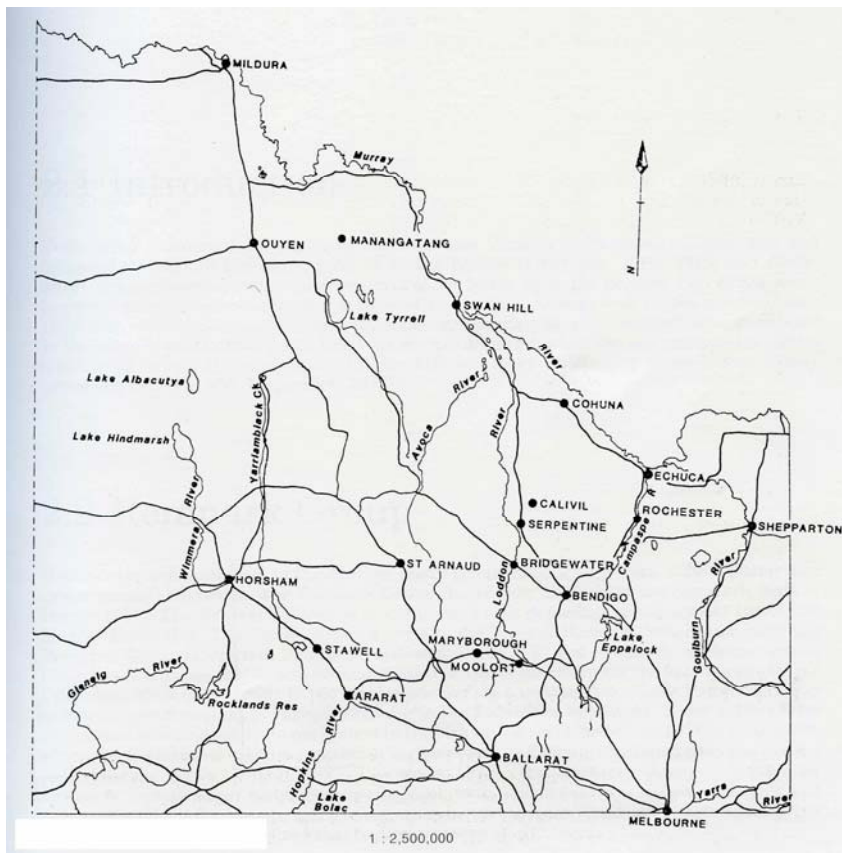


Figure 1.2 - Locality map of study area

2. GEOLOGY

2.1 Introduction

Continental sedimentation began in northwestern Victoria in Palaeocene times and had extended throughout the Murray Basin by the Eocene (Lawrence, 1966, 1975; and Abele, 1976). It has persisted in the eastern parts of the basin up to the present day. In the west, however, paralic sedimentation commenced in the late Eocene and continued until late Oligocene, when the earliest marine sediments are recorded. Marine sedimentation continued in the western parts of the basin until the middle Miocene, when the sea retreated from the basin prior to transgressing again in late Miocene to early Pliocene times before finally retreating (Abele, 1976; Macumber, 1978b).

Continental sedimentation in the Murray Basin commenced with a thick suite of sands and carbonaceous sediments - the Renmark Group (Lawrence, 1975; from the Renmark Beds of Harris 1971). The Renmark Group is a widespread unit deposited across almost the entire Murray Basin (Figure 2.1). In the western parts of the Murray Basin, in South Australia, the Renmark Group is overlain by marine sediments (Figure 2.2) and ranges from Palaeocene to Oligocene (Ludbrook, 1973); however, in western Victoria the range maybe from Eocene to late Oligocene (Abele et al., 1976). In the eastern parts of the basin (see below), the Renmark Group sediments are deposited in valleys incised into the Palaeozoic basement. Those valleys form the earliest coordinated drainage system in the Murray Basin (Macumber, 1978b). Conditions which were conducive to the deposition of carbonaceous sediments continued from the Eocene until Miocene times (Martin, 1977; Lawrence, 1975). The Renmark Group is divided by Lawrence into an upper carbonaceous silty unit, with additional clays, claystones, coals and sands - the Olney Formation, and a lower, medium- to coarse-grained sand unit - the Warina Sand. The Warina Sand is the basal regional aquifer in the Murray Basin.

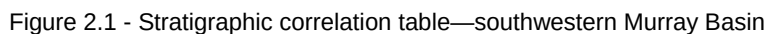


Figure 2.1 - Stratigraphic correlation table—southwestern Murray Basin

2.2.1 Renmark Group in the Eastern Murray Basin

2.2.1.1 Early Tertiary drainage systems

Structure contours of the pre-Tertiary surface under the southern Riverine Plain in Victoria show that a number of northwards draining valleys, with a drainage pattern similar to that existing at present, were in existence in early Tertiary times (Macumber, 1978b). These early Tertiary valleys can be traced downbasin from the point where the modern river valleys emerge from the highland front to a position beyond the present-day Murray River (Figure 2.3). The position of the valleys indicates that in Victoria the basic physiographic divisions and fluvial provinces were established before Eocene times and have undergone relatively little modification since that time. This is consistent with the view of Hills (1934, 1976), who considered that the uplift of the Eastern Highlands had largely occurred before early Tertiary times. It also supports the work of Wellman (1974), who considered that by the Oligocene the highlands were already in existence and possessed a relief of 1000 m, with drainage system similar to that of the present day. It seems likely that the basal valleys of the Riverine Plain are the downstream extensions of the early Tertiary highland valleys recorded by Wellman (1974) and Hills (1975).

The well-defined character of the tributaries suggests that at least a partly coordinated drainage system and therefore, an ancestral Murray River existed in the eastern Murray Basin in early Tertiary times. While it is not possible to be explicit about the trunk system, it is notable that the trends of the Campaspe and Goulburn valleys indicate a junction in the vicinity of Mathoura on the edge of the Cadell Tilt Block. This area was a focal point for later phases of Murray River activity in the late Quaternary and the Holocene periods (Pels, 1966; Bowler, 1967). It is not known to what extent the basement valleys are erosional and to what extent structural; however, it is probable that both factors have been variously operative. For instance, Tickell (1977) notes that the present-day distribution of Permian rocks southward into the Campaspe, Goulburn, and Ovens-King valleys suggests that the positions of the valleys were determined to a large degree by the same structures which preserved the Permian rocks. Similarly, in the Loddon Valley, the confinement of Permian glacial sediments to a narrow band within the highland tract of the Tertiary Loddon River (Figure 2.4) has led to the conclusion (Macumber, 1978d) that faulting played a significant role in the development of the Loddon Valley in Tertiary times.

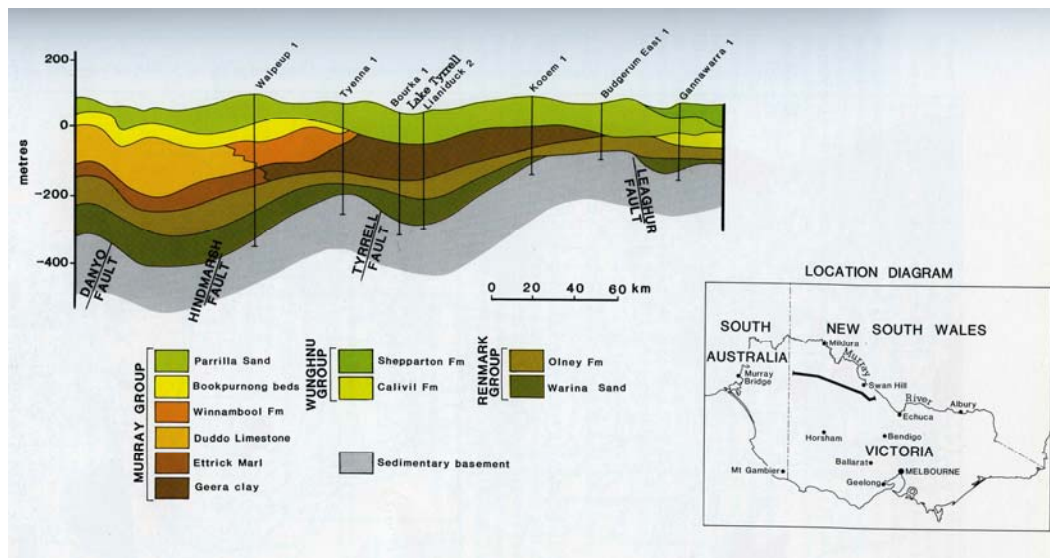


Figure 2.2 - East-west geological cross section through the southwestern part of the Murray Basin (modified from Lawrence, 1975)

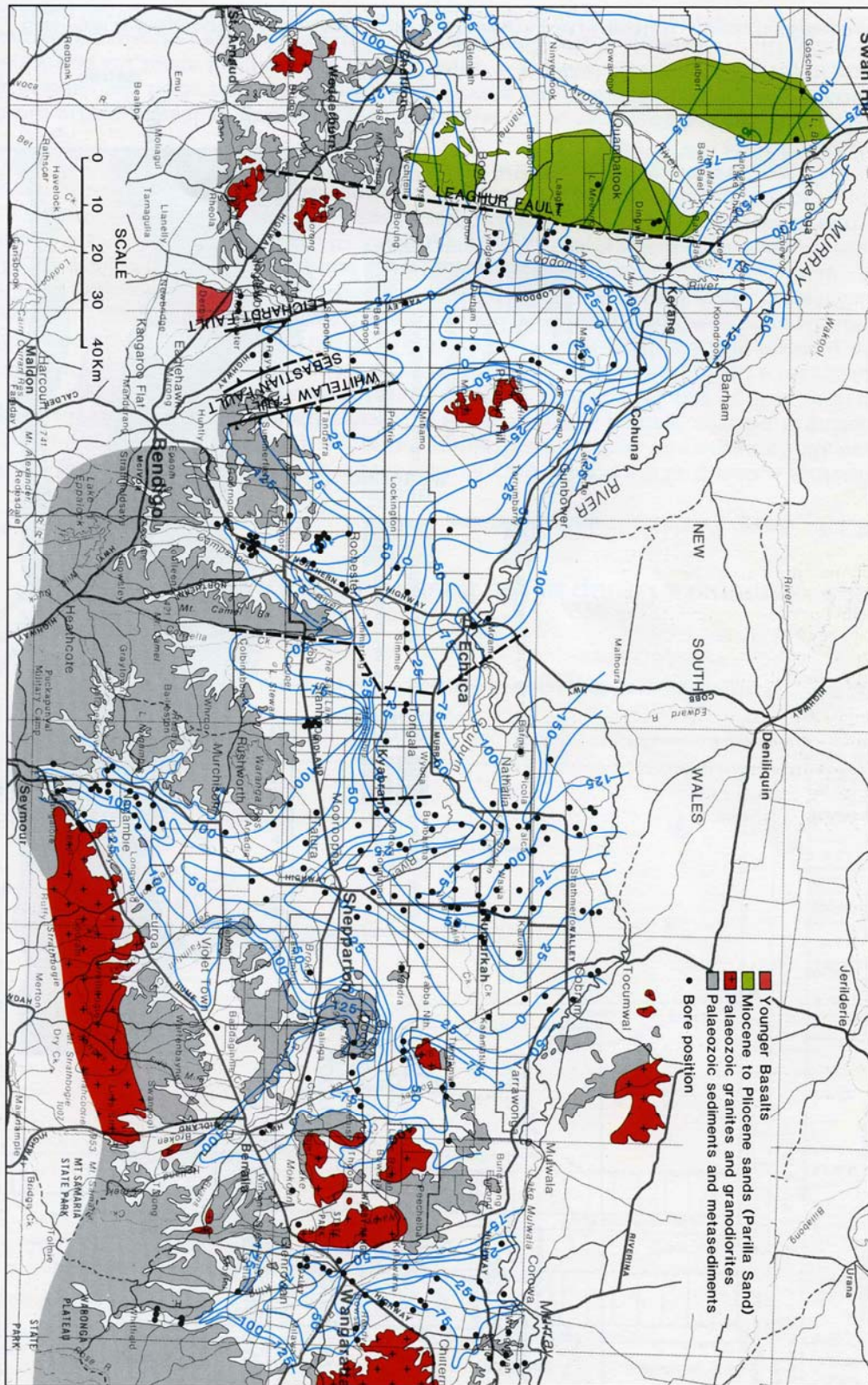


Figure 2.3 – Structure contours of the top of the Palaeozoic basement beneath the southern Riverine Plain (from Macumber, 1978b).

The significance of structural control in the continued development and deepening of the basement depressions is well illustrated in the case of the lower Loddon Valley where there is a concordance of levels between the uppermost surface of the Tertiary Renmark Group sediments, deposited within a slowly subsiding graben lying to the east of the Leaghur Fault (Macumber, 1969), and the adjacent stable Palaeozoic surface. This led to the development of an extensive plain - the Mologa Surface (Macumber, 1978a) - formed across both Tertiary sediments and peneplained Palaeozoic sediments. The Mologa Surface, now buried beneath later deposited sediments, covered all of the central Loddon Plains.

2.2.1.2 Backfilling of the valleys - the Renmark Group sediments

Sedimentation in the valleys began with deposition of a predominantly quartz sand lithofacies (Warina Sand, Lawrence, 1975), although gravel, ligneous clays and silts, and minor lignites are also present. Lawrence notes that the primary control appears to have been the drainage pattern at the time the Warina Sand was deposited. In some instances, the Renmark Group sediments can be traced up into the highland tracts of the valleys. Tickell (1977) notes that ligneous clay and gravel which are encountered in the Campaspe and Goulburn valleys can be traced down the valleys and are thought to be continuous with the Renmark Group. The picture is not so clear in the Loddon Valley immediately upstream of the highland front, where a late Tertiary valley incision has removed much of the Renmark Group sediments that might previously have existed within the highland tract. Instead, the valley is backfilled with coarse-grained quartz gravel, and pebbles which form the fluvialite "deep lead" facies of the Upper Miocene Calivil Formation (Macumber, 1973).

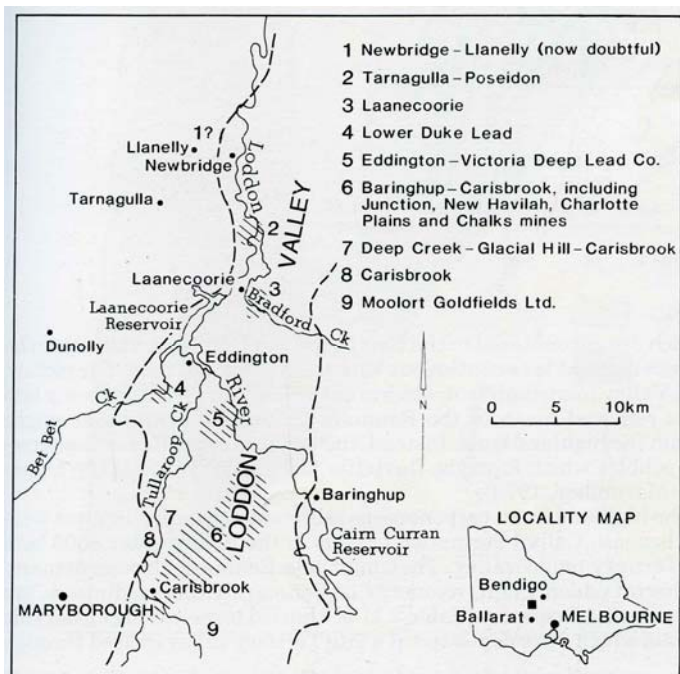


Figure 2.4 - Distribution of the Permian glacial beds in the Loddon River drainage basin is restricted to that area occupied by the Loddon Valley and its deep lead system (Macumber, 1978b)

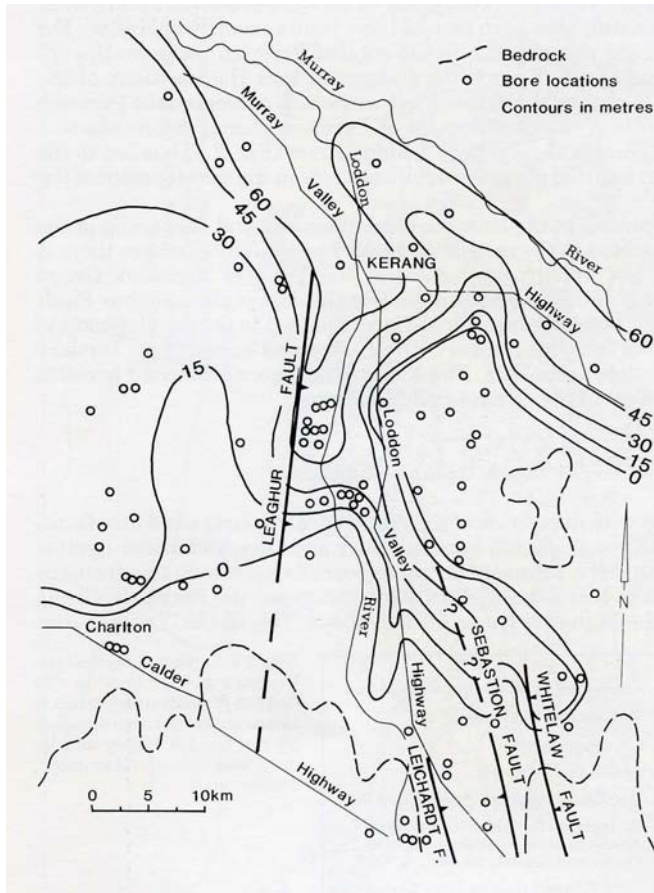


Figure 2.5 - Isopach map of the Renmark Group in the Loddon Valley (after Macumber, 1978b)

At Bridgewater close to the highland front, carbonaceous sediments including lignites were encountered at some depth beneath Calivil Formation gravels in the Bridgewater 8008 bore on the edge of the earliest Tertiary palaeovalley. They may be a Renmark Group remnant. Further north, on the southern Loddon Plain, remnants of Renmark Group sediments are encountered in bores with increasing frequency (Table 2.1) as a buried terrace lying either side of Calivil Formation sediments which were deposited in a late Tertiary valley incised through the Renmark Group. In the upper central Loddon Plain, the valley incision was not complete, and the Renmark Group underlies the Calivil Formation.

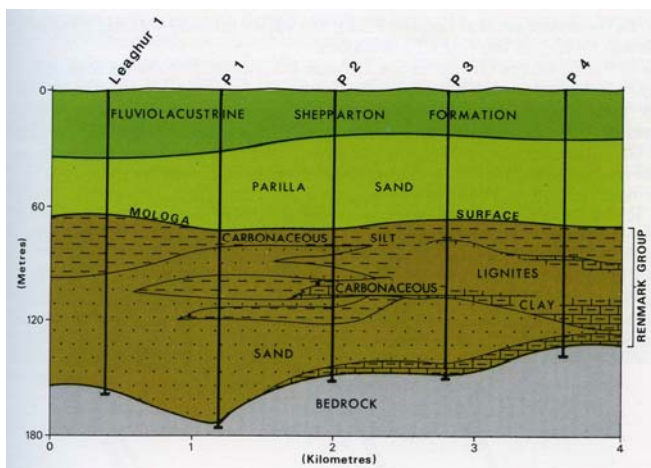


Figure 2.6 – East-west geological cross section through the Loddon Plain at Leaghur shows the Parilla Sand directly overlying the Renmark Group

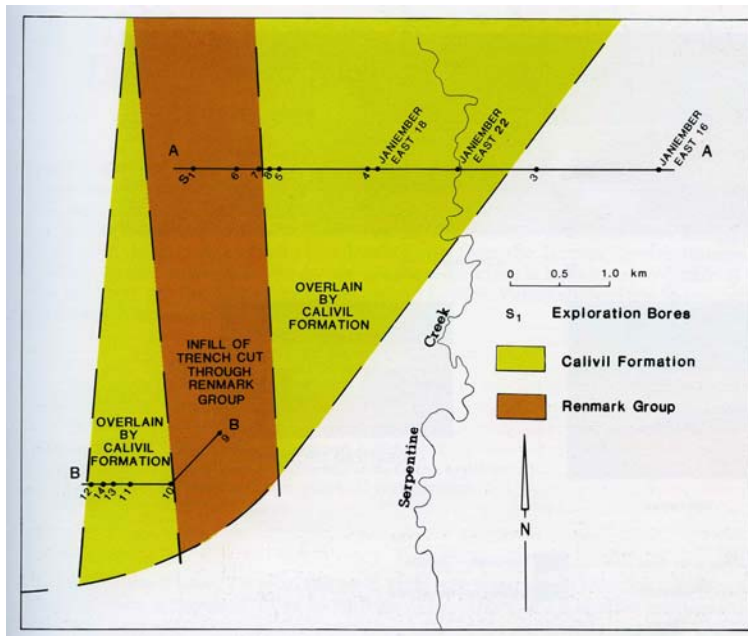


Figure 2.7 – The distribution of the Renmark Group and the Calivil Formation in the Powlett Plains palaeovalley, northwest of Serpentine (after McKenna et al., 1978).

The isopach map of the Renmark Group on the Loddon Plain (Figure 2.5) shows that only a thin veneer of the carbonaceous sediments occurs to the west of the Leaghur Fault, at the western edge of the plain. Beneath the western part of the plain, the Renmark Group sediments occupy a valley incised into Palaeozoic sediments. In the eastern parts of the Loddon Plain, the Renmark Group sediments are absent, and late Tertiary sequences sit directly on Palaeozoic sediments and metasediments. A closely spaced series of bores, drilled across the western Loddon Plains at Leaghur, shows the lithological variation in the Renmark Group sediments (Figure 2.6). It is notable that coarse quartz gravel and pebbles, so typical of the Calivil Formation, are not a prominent component of the sediment infilling the older Tertiary valley and are usually absent from the sequences in the Loddon Basin.

A clear picture of these two phases of valley incision and backfilling in early and late Tertiary times is seen in the Powlett Plains deep lead, a minor palaeovalley, now totally infilled, paralleling the also infilled palaeo-Loddon valley (Figure 2.7). Intensive drilling of the Powlett Plains palaeovalley provided a clear section of the valley fill (McKenna et al., 1978).

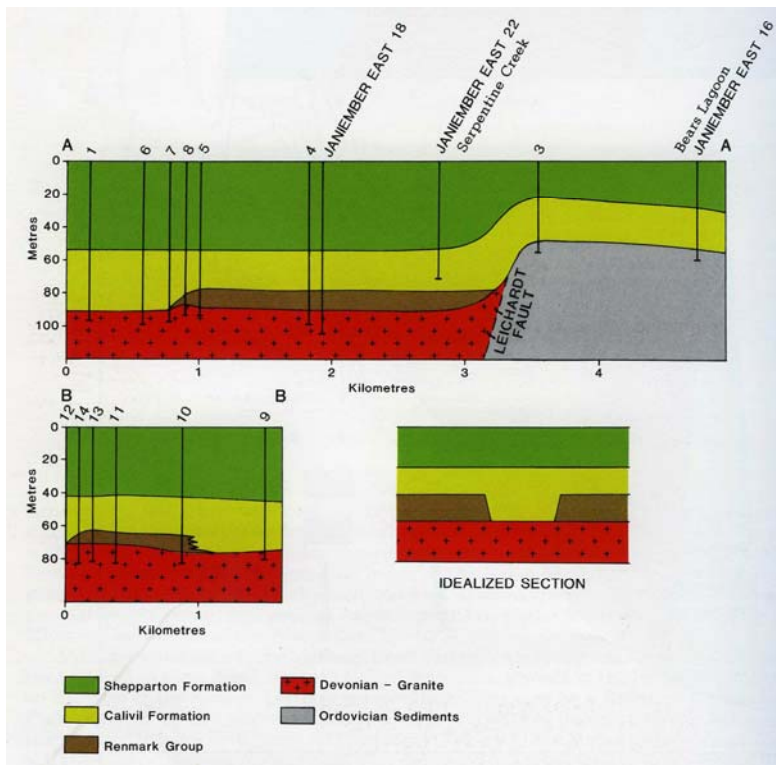


Figure 2.8 - Geological cross section through the Powlett Plains lead, to the west of Bears Lagoon

The earliest valley, cut into deeply weathered granite, was partially backfilled by medium- to coarse-grained quartz sands, which were sparingly carbonaceous and pyritic with interbedded thin, ligneous clay seams - the Warina Sand of the Renmark Group. A second phase of valley incision has cut through the Renmark Group sediments, leaving the remnants as a buried terrace along the sides of the newly rejuvenated valley. Subsequent partial backfilling was by coarse-grained sand and gravel of the Calivil Formation (Figure 2.8).

Throughout most of the Murray Basin away from the highland tract the Warina Sand is overlain by the Olney Formation (Lawrence, 1975), consisting of carbonaceous clay, sand, and lignite. Lawrence concludes that in the western part of the basin these sediments were deposited in deltaic, lagoonal, and tidal flat environments. However, in the absence of either marine or paralic conditions in the eastern part of the basin, it is concluded that lacustrine and paludal conditions were dominant. Martin (1977) puts the upper limits of carbonaceous sedimentation in the eastern part of the Murray Basin as Middle Miocene. The presence of lagoonal environments close to the highland front indicates a major shift in stream regime, whereby the earliest coordinated drainage of the Eocene system was replaced by uncoordinated drainage in the Oligocene. These conditions were maintained until middle Miocene times. During this interval, the Riverine Plain in Victoria is seen as being essentially a huge swamp.

2.3 Mologa Surface and Late Tertiary Tectonism on the Loddon Plain

Following the cessation of Renmark Group sedimentation in middle Miocene times, an extensive peneplain, the Mologa Surface (Macumber, 1978b), existed across the Loddon Plain. In the west of the plain, the surface is formed on carbonaceous Renmark Group sediments while in the east, it cuts across deeply weathered Ordovician sediments and metasediments (Figure 2.9) and the granites which extend westwards from the Terrick Terrick Ranges. Where formed across the Ordovician, the deeply weathered surface is kaolinized to depths of up to 30 m. For instance, the Loddon 1 bore in the central Loddon Valley showed the following bedrock (Ordovician) lithologies:

LODDON 1

Depth (m)	Lithology
(0.0 - 67.0	Late Tertiary to recent sediments)
67.0 - 91.5	White clay (weathered Ordovician sediments)
91.5 -105.5	White clay with quartz (? reef quartz)
105.5+	Grey mudstone

In the Loddon 2 bore, 7.4 m to the east, similarly weathered Ordovician sediments were intersected from a depth of 70 m to 94.5 m. While there is no similar weathering on the Renmark Group sediments, it does occur on Ordovician sediments beneath the Renmark Group, suggesting that it correlates with that phase of weathering associated with the Mesozoic palaeoplain observed by Hills (1940). There is no evidence of ferruginization of the Mologa Surface, an event which only appears on the Loddon Plain as a significant feature in sediments deposited well after the deposition, of the Renmark Group; ferruginization is most prominently developed in the late Tertiary marine Parilla Sand sequence. The Mologa Surface was later covered by fluviolacustrine and marine deposits and is now met only in bores at depths from about 67-78 m below the present surface. A list of bores where the Mologa Surface is clearly identified is given in Table 2.1. On passing northward across the Loddon Plain, the surface has an overall slope of about 1.2 m/km which is much the same as the present-day slope of the plain. Some minor differences between these two surfaces are seen in the slightly greater depth to the Mologa Surface in the southern parishes. However, still further south again, the Mologa Surface rises to be at a depth of 61 m at Bridgewater.

TABLE 2.1 - Characteristics of the Mologa Surface-Loddon Plain

	Ground-surface Elevation (m-AHD)	Depth to Mologa Surface (m)	Elevation of Mologa Surface (m-AHD)	Stratigraphic unit on which the Mologa Surface developed
Tandarra 1	105.5	73.2	32.0	Renmark Group
Janiember East 10001	106.7	78.4	28.3	
Jarklan 8001	102.1	76.5	25.6	
Hayanmi 8001	102.1	78.4	23.7	
Leaghur (P) (average 4 bores)	86.9	64.0	22.9	Ordovician Sed.
Yarrowalla 8001	93.9	71.6	22.3	
Boort 24	88.4	67.7	20.7	
Loddon 1	83.8	70.1	19.7	
Boort 22	87.8	70.1	17.7	Renmark Group
Leaghur 2	86.3	68.6	17.7	
Leaghur 1	87.2	70.1	17.1	
Leaghur (P) (3 bores)	84.1	67.0	17.1	
Leaghur (P) (4 bores)	86.0	70.1	15.9	Ordovician Sed.
Mincha West 10	84.1	70.1	14.0	
Meran 10	79.3	75.0	4.3	

P= Private bores drilled for uranium

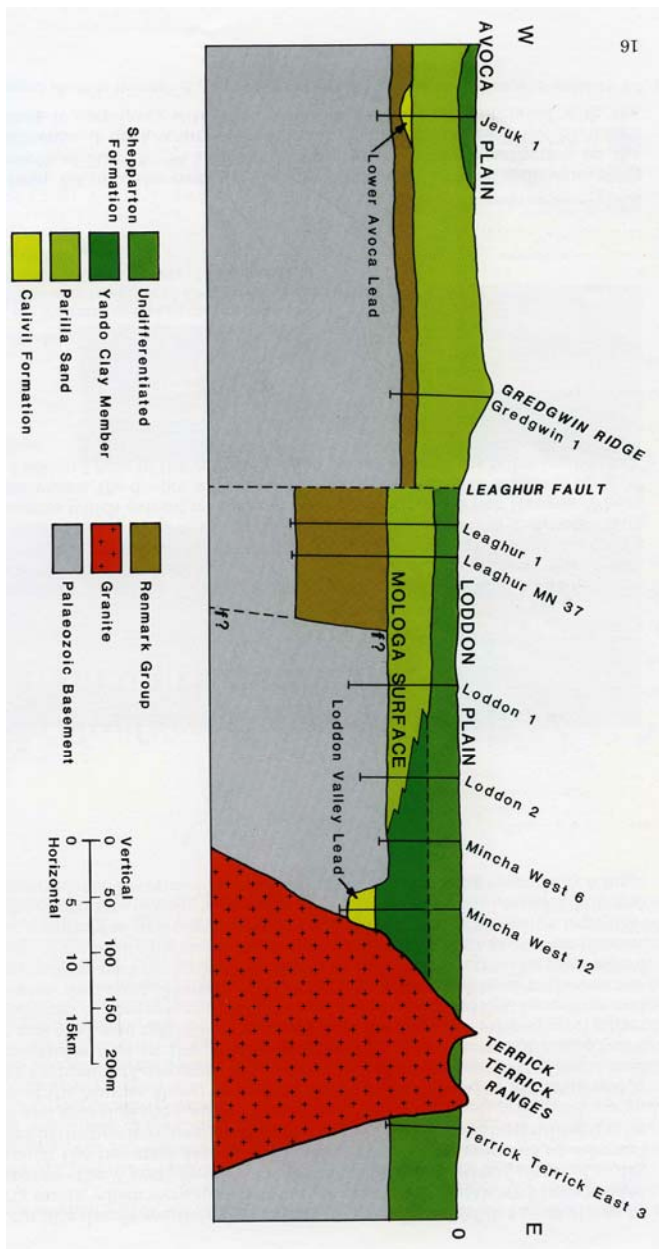


Figure 2.9 – Geological cross section through the central Loddon and Avoca Plains.

While the Mologa Surface is well known on the Loddon Plain, bore hole data from east of the Terrick Terrick Ranges, at the limit of the Loddon Plain, suggest that it extended eastward across the Campaspe Plain. That is, in middle Miocene times, the granitic Terrick Terrick Ranges protruded from a weathered surface cut across Ordovician sediments. To the east of the range, the deeply weathered Mologa Surface was met at a depth of about 68.5 m in the Terrick Terrick 1 bore which is about the same depth as it occurs beneath the Loddon Plain. In order to explain the concordance of levels of the Palaeozoic surface and Renmark Group across the Mologa Surface, the Renmark Group is envisaged as having been deposited in a slowly subsiding graben structure, the Yando Trough, formed adjacent to a pre-existing bedrock surface, itself formed as a pediment to the Terrick Terrick Ranges (Figure 2.9). Ongoing subsidence of this structure is seen as controlling the overall position of the early and middle Tertiary drainage system on the Loddon Plain in much the same way as occurred within the highland tract, where the late Tertiary Loddon Valley drainage pattern has been shown to have been ultimately controlled by a tectonic influence (Macumber, 1978d). An examination of the late Quaternary stream traces of the Loddon River shows that the various sheet flood systems of distributaries, anabranches, and playettes (Macumber, 1968) on the Loddon system (Figure 2.10) in the central Loddon Plain are virtually restricted to the graben. Furthermore, the many lunette-lake systems, from Boort to beyond Kerang, all lie within the trough.

The controlling faults bordering the graben are the Leaghur Fault to the west and the northern continuation of the Sebastian Fault to the east. The Sebastian Fault has been shown by Whitelaw (1899) to have been active in Pliocene times, when it downfaulted the Sebastian deep lead (Calivil Formation) at Sebastian. Further evidence of the influence of tectonics on Tertiary sedimentation in the southern Loddon Plain can be seen in the Dingee area, where the Tandarra 1 bore passed through over 60 m of Renmark Group sediments before reaching basement at 135 m. The Dingee 1 bore, to the east, bottomed at 54 m; the Talambe 1 bore, to the west, bottomed at 57 m; and the Tandarra 2 bore, to the south, bottomed at 37 m. Renmark Group sediments were absent from all three bores. The presence of the low at Dingee may be related to Tertiary movements on the Whitelaw Fault (Figure 2.5), the northern extensions of which pass immediately to the east of the Talambe 1 bore, and, therefore, to the west of Dingee.

More recent tectonic activity along the Whitelaw Fault may account for the uncharacteristic easterly swing of Myers Creek beyond Raywood into the Dingee area, from its initial northerly flow direction following the general fall of the plain. Late Pleistocene stream systems including distributaries of the Loddon River and Pyramid Creek show a similar trend.

2.4 Murray Group

In the western parts of the Murray Basin, the Renmark Group is overlain by a marine sequence referred to as the Murray Group (Ludbrook, 1957). Lawrence (1966) extended the concept of the Murray Group to include the Ettrick Marl (Ettrick Formation of Ludbrook, 1957). The Murray Group of Lawrence includes all the Tertiary marine units overlying the Renmark Group in the Murray Basin, with the exception of the uppermost two units - the Bookpurnong Beds and the Parilla Sand. It includes the Ettrick Marl, Geera Clay, Duddo Limestone, and Winnambool Formation and ranges from late Oligocene to middle Miocene in age (Figure 2.1). During the initial stages of marine sedimentation, a grey fossiliferous marl (the Ettrick Marl) was deposited in the western parts of the basin, and this grades eastward into a sparingly fossiliferous dark grey to black clay - the Geera Clay (Lawrence, 1966).

Overlying the Ettrick Marl, conformably and diachronously, is the Duddo Limestone (Lawrence, 1966), which attains a thickness of 130 m in northwestern Victoria. This is a buff-coloured, bryozoal calcarenite, with interbedded layers of calcisiltite and calcilutite. The high permeabilities and good quality water found in the Duddo Limestone render this unit a major aquifer for stock and domestic water supplies. The Duddo Limestone grades laterally eastward into a glauconitic, marly clay or marl, and sometimes a limestone; it contains a largely molluscan fauna, and, in turn, grades eastward and downward into the Geera Clay, with which it has a diachronous boundary.



Figure 2.10 – Sheet flood landscape developed on the western Loddon Plain

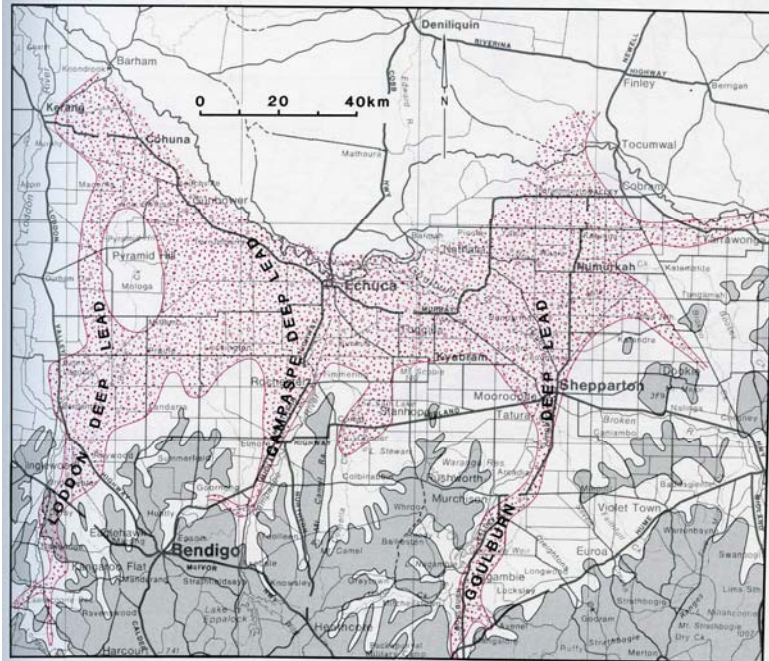


Figure 2.11 – Generalized distribution of the late Tertiary deep lead sediments beneath the southeastern Riverine Plain in Victoria (from Macumber, 1978a).

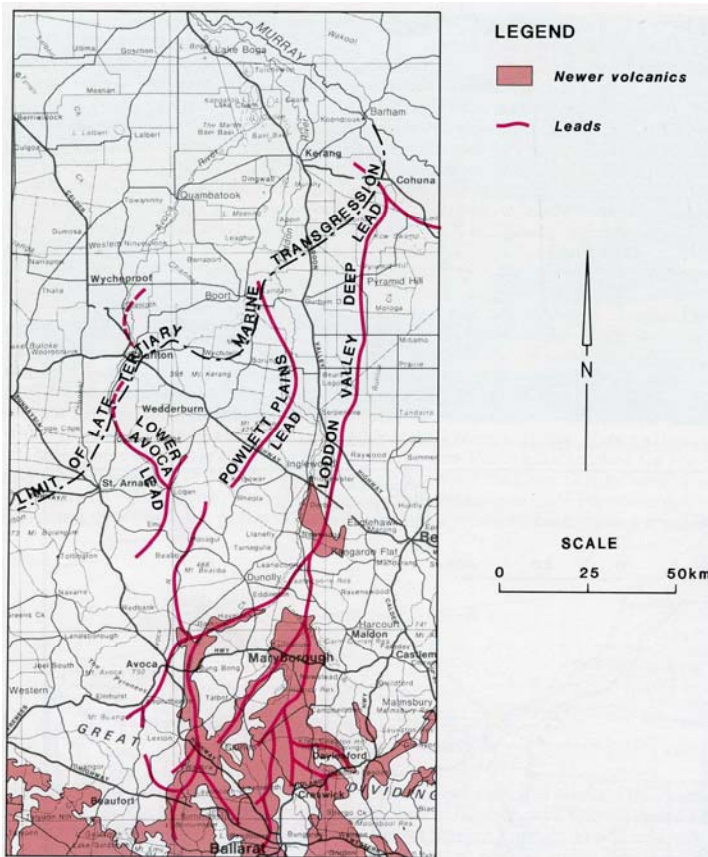


Figure 2.12 – The Loddon Valley, Avoca Valley and Powlett Plains drainage systems in late Tertiary times.

2.5 Calivil Formation

Following the deposition of the Murray Group, the sea regressed for a short period in the middle Miocene (Lawrence and Abele, 1976), prior to a final transgressive cycle in the late Miocene to early Pliocene. This intervening period during the regression is seen as one when the later of two major phases of valley incision occurred in northern Victoria in response to a general lowering of base levels. This reestablished, in late Tertiary times, a coordinated Murray River drainage system along the lines of that which existed in the Eocene (Macumber, 1978b). The physiographic divisions of the Murray Basin which first appeared in the early Tertiary were firmly established by late Tertiary times; these systems have undergone only minor modification since that time.

The transgression in the late Miocene again raised base levels, causing a gradual backfilling of the valleys by fluvial sand and gravel of the Calivil Formation (Macumber, 1973). At the inland limits of the marine transgression in the east (beneath the Loddon Plain), the Calivil Formation is overlain by and laterally grades upward into the marine Parilla Sand (Macumber, 1978b); the boundary between the Calivil Formation and the Parilla Sand is therefore diachronous.

Within the highlands, the Calivil Formation sediments are confined to major valleys incised into Palaeozoic basement. However, on passing across the Riverine Plain, the river systems were less confined and fanned out. Late Tertiary sand and gravel are now found as a sheet deposit underlying the lower Campaspe and Goulburn plains (Figure 2.11). In the far northern areas adjacent to the present-day Murray River, the individual tributary fluvial provinces merge into a trunk Murray Valley province. However, because of the migratory character of the systems this changeover point is blurred, and therefore an arbitrary line is taken for the commencement of the Murray Valley province, which corresponds to the zone where individual south-north sand tongues merge to make a continuous east-west sand sheet.

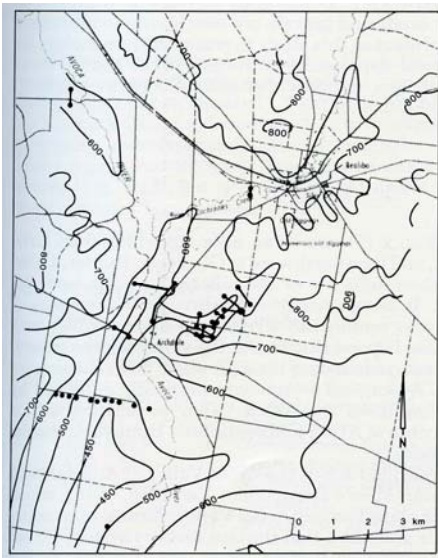


Figure 2.13 - Structure contours of the base of the Avoca Valley near Archdale, showing the position of the Tertiary stream divide as being upstream of the junction of Cochrane's Creek and the present-day Avoca River, which now flows north (contours in feet)

In order to gain some understanding of the stratigraphic relationships between the late Tertiary marine and non-marine sequences in the Murray Basin, two major late Tertiary river systems (the Loddon River and Avoca River) and a third smaller system (the Powlett Plains deep lead)

were examined in detail, and their respective sedimentary sequences were traced downbasin to a point where they could be related to marine sequences of the western Murray Basin (Figure 2.12).

2.5.1 Palaeogeography of the Avoca Valley and Loddon Valley Leads - General

The Loddon Valley and Avoca Valley deep leads are late Tertiary drainage systems passing northward from an east-west Tertiary divide situated a little to the south of the present main east-west divide (Thomas, 1952). The position and shape of the Tertiary valleys within the highlands is well documented as a result of detailed work during the gold mining era (Hunter, 1909); however, the investigations stopped far short of the highland front, thus giving no clue to their downstream continuation beneath the plains.

The river valleys were called 'deep leads', and this term is widely used in the gold mining literature, for instance, Hunter's (1909) classic work, *Deep Leads of Victoria*. Although on passing plainwards the valley-fill deep lead sands and gravels are now known to ultimately become sheet sand deposits, it has been convenient in this study to retain the terminology for these downstream extensions. The term is most appropriate in the case of the Loddon River valley, which retains its incised valley morphology virtually throughout its entire length.

2.5.2 Avoca Valley Lead

The headwaters of the Avoca Valley lead (Figure 2.12) rise in an area stretching from Lake Burrumbeet through the Beaufort, to Raglan, and then northward to Glenlogie (Hunter, 1909). The system gains trunk proportions after the confluence of the Lillerie-Trawalla lead, the Ballarat West lead, and the Beaufort lead. It passes northwards through the parishes of Caralulup, Lillicur, and Glenmona, where it is confined to a relatively narrow valley now occupied by the Caralulup and Bet Bet creeks. Beyond Glenmona, the valley widens considerably to form the Nattyallock - Archdale Basin, northward through which flows the present Avoca River. The course of the main trunk Avoca lead system was originally described by Murray (1898) and Hunter (1909) as passing down the Avoca Valley to junction with a tributary system commencing near the township of Avoca and continuing through Archdale, Nattyallock, and Tunstals towards St. Arnaud.

Some time after this, a series of bores was drilled across the Avoca Valley at Archdale and near Bealiba (Annual Reports of the Victorian Mine's Department Vic. 1915, 1916), which provided new information on the position of the late Tertiary Avoca Valley. Structure contours of the base of the Avoca Valley in the Archdale area show that the late Tertiary system did not flow northward towards the Avoca River plains as it does today, but, instead, flowed southward, towards Nattyallock, where it became distributary to a northward flowing system (Figure 2.13). The enlarged river system then turned eastward down the Bet Bet Valley to junction with the late Tertiary Loddon Valley system at Eddington.

During late Tertiary times a stream divide existed in the vicinity of Tunstals, where the Avoca Valley today has only 26 m of alluvial fill. The elevation of the base of the divide at Tunstals is 161 m, while at Eddington it is 108 m. North of the Tunstals divide, a relatively minor lower Avoca system commences and passes downbasin, through Logan and Gower, where the valley is only about 2.5 km wide. This situation continues to Coonoor Bridge, where the valley leaves the highland tract. Beyond this point, the course of the Avoca lead passes out under the Avoca Plain and there underlies marine sediments (the Parilla Sand), deposited during the late Tertiary (Upper Miocene to Lower Pliocene) transgression. The transgression overlapped the highlands, in the vicinity of Coonoor Bridge and passed eastward to Boort. At Doboobetic to the west of Coonoor Bridge, in the Doboobetic 1 bore, the Parilla Sand occurs to a depth of 36.6 m and overlies Renmark Group sediments; there is no evidence of the Calivil Formation.

DOBOOBETIC 1

Depth (m)	Lithology
6.0 - 24.4	
24.4 - 27.4	Red-brown, silty, fine-grained sandstone
27.4 - 30.5	Red-brown, silty clay to fine siltstone
30.5 - 33.5	Light grey sandstone
33.5 - 36.6	Light greyish brown silty sandstone
36.6 - 61.0	Carbonaceous silts and sands - Renmark Group
61.0 +	Weathered Palaeozoic rocks (basement)

While there was no sign of Calivil Formation gravel in the Doboobetic bore, further to the west coarse gravel with quartz pebbles up to 25 mm diameter was obtained from the Teddywaddy 1 bore, situated 7.2 km north of Charlton (for location, see Figure 2.17). There were no Renmark Group sediments in this bore, suggesting that a late Tertiary valley had incised previously underlying carbonaceous sediments and totally removed them. The position of the bore suggests that the lower Avoca course was much the same as the present-day Avoca River course (Figure 2.12), but it is now buried under Parilla Sand sediments. A summary of the Teddywaddy 1 bore follows:

TEDDYWADDY 1

Depth (m)	Lithology
0.0 - 30.0	Reddish brown clay to sandy clay - Shepparton Formation
30.0 - 36.6	Light grey to white, fine, micaceous sand - Parilla Sand
36.6 - 45.7	Yellowish brown, fine- to medium-grained sand
45.7 - 48.7	Strongly ferruginized, micaceous, medium- to coarse - grained sand
48.7 - 57.9	Yellow brown, fine- to medium-grained sand
57.9 - 64.0	Coarse-grained sand and gravel - Calivil Formation
64.0 - 68.6	Well-rounded, coarse gravel with pebbles up to 25 mm diameter
68.6 +	Weathered Palaeozoic rocks (basement)

It is notable that in the adjoining parish of Narrewillock to the east, fossiliferous Parilla Sand outcrops about 4 km from the Avoca River. There are few deep bores beyond Teddywaddy until the Jeruk 1 bore at Jeruk about 17.5 km further north. This bore has a similar sequence, with fine micaceous sand (Parilla Sand) overlying fine- to coarse-grained sand and gravel, which occupy the interval from 55 m to 75 m. Beneath the gravel, the sequence passes downwards into the lignite and ligneous clay of the Renmark Group.

JERUK 1

Depth (m)	Lithology
0.0 - 15.2	Sample missing- Shepparton Formation
15.2 - 29.0	Clay and sandy clay
29.0 - 54.9	Fine sand and clay - Parilla Sand
54.9 - 70.1	Fine- to coarse-grained sand with occasional pebbles
70.1 - 74.7	Calivil Formation
	Coarse-grained sand to gravel
74.7 - 85.4	Carbonaceous silt, clay, and lignite - Renmark Group
85.4 +	Weathered Palaeozoic rock (basement)

The Jeruk 2 bore, drilled at Glenloth East, a further 4 km to the north, contained gravel in a similar stratigraphic position to that of Jeruk 1; however, this was the most northerly point at which Calivil Formation sediments were found in the Avoca Valley. Northward beyond Jeruk, no evidence has yet been found for any continuation of the lead; instead, the interval which the lead occupies farther south is occupied by marine sediments. For instance, in the Quambatook 1 bore, situated

on the Avoca River at Quambatook, fossiliferous micaceous sandy silts occur from 53 m to 75 m, and these overlie ligneous silts (Macumber, 1969). To the south of Quambatook, in the parishes of Ninyeunook and Towaninny, eight bores show sparingly fossiliferous fine-grained sandstones (the Parilla Sand) overlying glauconitic pyritic sands and silts, which in turn overlie carbonaceous sediments. Here, the glauconitic unit occupies the interval from about 60 m to 100 m, normally occupied by the Calivil Formation. Since there is no palaeontological data on the glauconitic interval, there is no control on its age; however, its lithology is closer to the Bookpurnong Beds in character than to the somewhat older Geera Clay (see stratigraphic table - Figure 2.1). The issue is complicated by the presence of marine sediment at Wycheproof, about 18 km to the west of Jeruk. These sediments are more readily correlated on lithology with the Geera Clay, but, again, there is no palaeontological data to support this. This problem underlines the difficulties of making lithological correlations at the eastern limits of the Murravian transgression, where, in the absence of calcareous units such as the Duddo Limestone, many units are essentially sparingly fossiliferous, dark grey clay and sandy clay, which cannot be readily distinguished on lithology alone. The problem is exacerbated by the broadened, and at times overlapping lithological definitions, which are required to cater for the wide range of depositional environments represented by each unit.

On passing eastward, beyond the Avoca River, this stratigraphic problem no longer exists, since the Avoca River marks the approximate inland limits of the middle Tertiary transgression. Only the late Tertiary transgression passed beyond this point, and marine sequences east of the Avoca River are represented essentially by the Parilla Sand. While there is some doubt about the stratigraphic position of the glauconitic sediments in the Quambatook South bores, this is not the case in the Quambatook 1 bore at Quambatook, where fossiliferous silts are seen as being the initial transgressive unit in the late Tertiary transgression (Macumber, 1969). The absence of Calivil Formation lithologies in the Quambatook area, where their stratigraphic interval is instead occupied by marine sequences suggests that a shoreline has been crossed, with the fossiliferous silt there representing the marine equivalents of the fluvial Calivil Formation. Therefore, it seems likely that the late Tertiary lower Avoca River system flowed into the sea somewhere between Jeruk and Quambatook. A further pulse of marine transgression resulted in the fluvial sequence being covered by the Parilla Sand, with drowning of the Avoca Valley at Coonoor between Charlton and St. Arnaud (see Figure 2.12).

2.5.3 Powlett Plains Lead

A situation similar to that occurring on the Lower Avoca system occurs on the Powlett Plains Lead, which passes northward beneath the western Loddon plain towards Yando (Figure 2.12); this system parallels, but does not join the Loddon Valley system. Unlike the neighbouring Loddon Valley system, the Powlett Plains Lead commences locally, being fed by two tributaries - the Old Inglewood Lead and the Kingower Lead, which drain small adjacent catchments. After junctioning at Glen Albion, the tributaries form a single Powlett Plains system, which trends northward past Powlett into the Serpentine area, where it passes beneath the Loddon River into the Loddon Plain. The system continues northward towards Yando, separated from the Loddon Valley system by a buried north-south trending bedrock ridge, which is positioned about 1 km west of Bears Lagoon (Figure 2.7).

In the Serpentine area, the late Tertiary Powlett Plains system has cut a shallow trench into an earlier valley fill of Renmark Group sediments (McKenna et al., 1978) - these latter carbonaceous sediments remain as remnants along the valley sides (Figure 2.8). The width of incision is about 1 km, while the base of the incision is about 90 m below the present surface, that is, much the same depth as that occurring in the Loddon Valley system farther east. This indicates that both systems at the time of their incision were graded essentially to the same level. Calivil Formation sediments come to within 54 m and 43 m respectively of the surface in two section lines drilled for sedimentary uranium (Figure 2.7); they are about 23 m thick where they overlie the Renmark Group and 36 m thick where they backfill the buried valley, which is incised into the Renmark Group (McKenna et al., 1978).

The Calivil Formation is overlain by the Shepparton Formation (Lawrence, 1966), a suite of finer-grained sediments, consisting of clays with interbedded shoestring sand lenses. In the Serpentine area the contact between the Calivil Formation and the Shepparton Formation is poorly defined and is generally contained within a zone of ferruginization (?lateritization). The lead continued into the Yando area, where two closely spaced lines of bores (Figures 2.14 and 2.15) show typical coarse sand and gravel of the Calivil Formation overlying the Renmark Group, and, in turn, being overlain by fluviatile, red and grey mottled clay. A section further north, at Yarrowalla West contains no gravel and, instead, shows interbedded sands and clays (Figure 2.16). Further north again, the interval above the Renmark Group is occupied by a sequence of marine Parilla Sand sediments (Figure 2.6), and the Calivil Formation was not present in any of three lines of bores. As in the case of the Lower Avoca system, this abrupt change in lithologies was interpreted as the crossing of a late Tertiary shoreline. On the Gredgwin Ridge, immediately to the west of these sections, fossiliferous Parilla Sand outcrops in a number of quarries at Boort and Leaghur (Macumber, 1969).

A line of bores, drilled in order to relate the marine and nonmarine sequences (Figure 2.17), shows that between the wholly marine and fluvial sequences there is a zone where the Calivil Formation gravels are overlain by marine Parilla Sand (Figure 2.18). Separating the two units is a ferruginous clastic sequence, consisting of fine-grained yellow-brown micaceous sand-stone, and ferruginous white to light brown sandy clay. The ferruginous sediments overlie either the Calivil Formation gravels or the Mologa Surface, which there is developed on carbonaceous Renmark Group sediments. They are in turn overlain by either the Parilla Sand in the north, or by oxidized red and grey mottled clays in the south. In the Boort 25 bore the sequence is about 15 m thick, occupying the interval from 58 m to 73 m and there overlying the Calivil Formation and underlying the Parilla Sand. Inland of the shoreline, the ferruginous zone becomes less clayey and passes into a fine to medium sand with clay lenses. It is usually overlain by red and grey oxidized clay - the Yando Clay member of the Shepparton Formation (Chapter 2.7) - which is considered to be a lateral, nonmarine facies equivalent of the Parilla Sand. When found under the Parilla Sand, the ferruginous sequence appears to be a basal, elastic precursor to the marine sequence; however, landward of the shoreline, it better represents the final phase of coarse, clastic, nonmarine sedimentation on the plain and is therefore, grouped with the Calivil Formation. The sediments are seen to represent a facies which arose after backfilling of the incised lead trough. The subsequent development of a braided channel system dumped sand on a coastal plain ahead of the advancing sea. The sequence is therefore considered to be fluviatile in the east and south, but it becomes deltaic or even littoral to the west where it cuts out and where Parilla Sand directly overlies the Renmark Group.

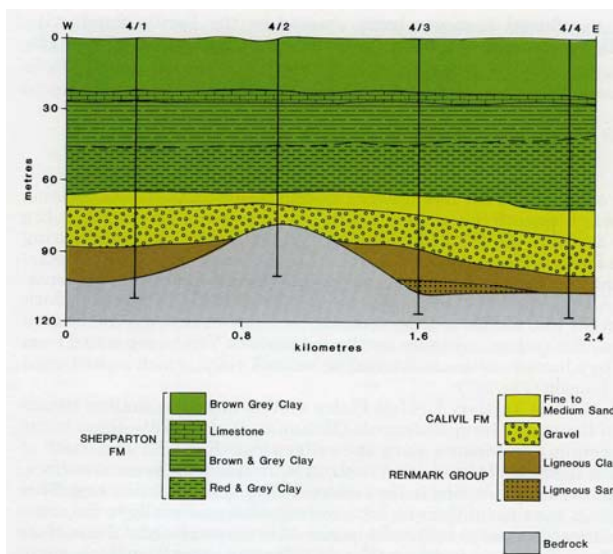


Figure 2.14 - Geological section through the western Loddon Plain at Yando (for section line, see Figure 2.33)

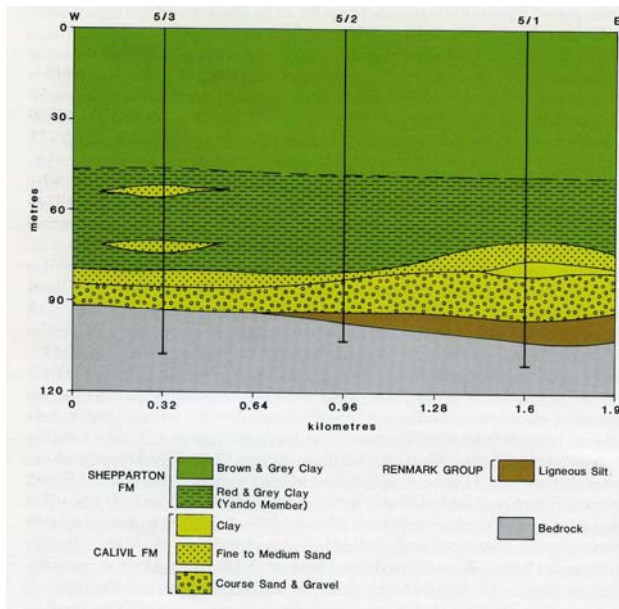
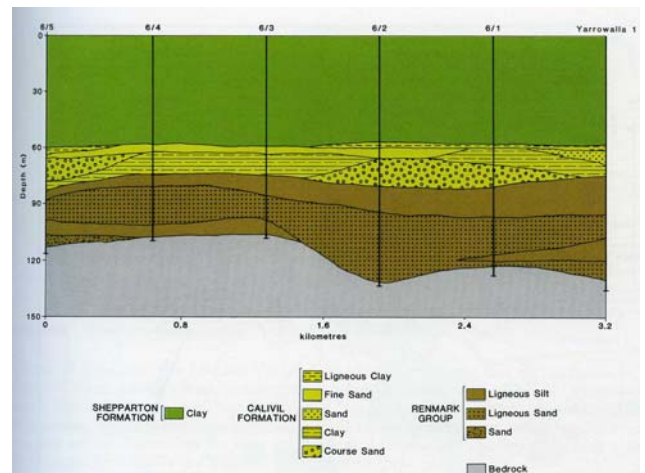


Figure 2.15 - Geological section through the western Loddon Plain at Yando

Figure 2.16 - East—west geological section through the western Loddon Plain at Yarrowalla



The ferruginized sediments are the first indication in the Tertiary sequences of the Loddon Plain of the onset of a phase of lateritization which continued to be active at the time of deposition of the overlying red and light grey mottled clays. The sequence at Yando is a repetition of that recorded at Serpentine, where the boundary between the Calivil Formation and the Shepparton Formation is contained within a ferruginous zone where red and grey clay overlies Calivil Formation gravel (McKenna et al., 1978). In northwestern Victoria lateritization is widely recorded from the Parilla Sand, where postdepositional weathering processes have produced a silicified and ferruginized capping (Figure 2.27) the Karoonda Surface of Firman (1966). Lawrence (1966) notes that the lateritized horizon developed on the Parilla Sand rises up into the Dundas Highlands. Both Firman and Lawrence regard the lateritized surface as Pliocene. A similar event is recorded on the late Tertiary sediments of the Otway Basin in southwestern Victoria, where the Miocene marine sediments are weathered and have produced a comparable lateritic soil, which is referred to by Gill (1964) as the Timboon Surface. Hills (1975) notes that the lateritic profile is well developed on a planated surface of Pliocene age formed along the

flanks of the Otway Ranges and well displayed to the east of Anglesea.

In the Loddon Basin, ferruginizing conditions first appear at the top of the Calivil Formation in the east and south of the Loddon Plain, and at the base of the Parilla Sand in the western areas. These conditions remained active during the deposition of the Yando Clay (the nonmarine lateral facies equivalent of the Parilla Sand) and continued for some time after the retreat of the sea in early Pliocene times. On stratigraphic evidence from the Loddon Plain, the age of this ferruginizing event ranges from about the Miocene - Pliocene boundary up to perhaps middle Pliocene times. As is shown in later chapters, the remobilization of iron oxide by groundwaters contributes strongly to the hydrochemical behaviour of the groundwater where it emerges in the discharge zones of northwestern Victoria.

2.5.4 Loddon Valley Lead

The ancient effluents tributary to the Loddon Valley system (Figure 2.12) commenced near Ballarat, Bullarook, Spring Hill, Creswick, and Lake Learmonth (Hunter, 1909). After the confluence of the Bullarook-Mount Prospect lead and the main Ascot - Clunes lead, in the parish of Smeaton, the Loddon Valley lead gains substance, and it continues as the trunk Berry-Moolort-Loddon lead, passing northward through the parishes of Glengower, Rodborough, Moolort, Baringhup, Neereman, and Laanecoorie. The course passes into the parish of Woodstock, which is the most northerly point where drilling was carried out during the gold-mining era to determine its true position and character. The early miners classified the sediment which infills the auriferous leads into three major types:

1. An upper clayey sequence of no interest for gold content.
2. "Drift" sand - fine to medium quartz sand, water saturated, and behaving like a "natural soup"; this material caused considerable problems during shaft sinking and driving.
3. The gold-bearing "wash" at the base of the sequence, described by Dunn (1905) as great well-rounded boulders and pebbles; some of the boulders range up to 2 or 3 tonnes in weight, and many of them are from "one to several hundredweight a piece." The most abundant size however is what Gregory (1903) aptly described as cannon shot: "the well rounded quartz pebbles that form the bulk of the wash dirt." Gravel may also be associated with the drift and/ or the wash. A typical section is seen in the Mines Department bore, Rodborough 20 (Annual Report Victorian Mines Department, 1900, p. 62), which passed through the following sequence:

RODBOROUGH 20

Depth (m)	Lithology
0.0 - 6.6	Surface clay
6.6 - 40.2	Basalt
40.2 - 69.2	Clay, occasionally sandy
69.2 - 111.0	Drift
111.0 - 114.0	Wash

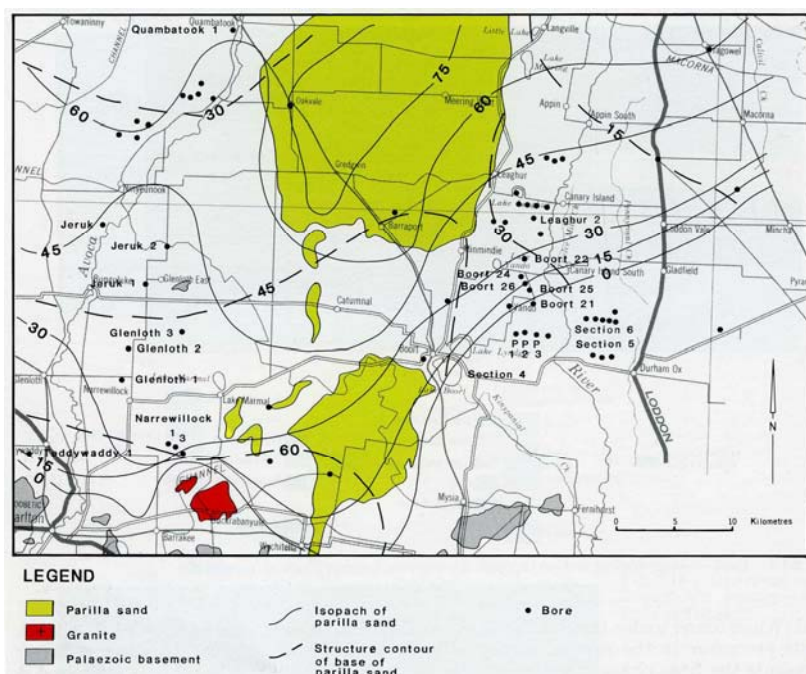


Figure 2.17 - Isopach map and structure contours of the Parilla Sand at the inland limits of the transgression. A section along the line between the Leaghur and Boort bores is given in Figure 2.18

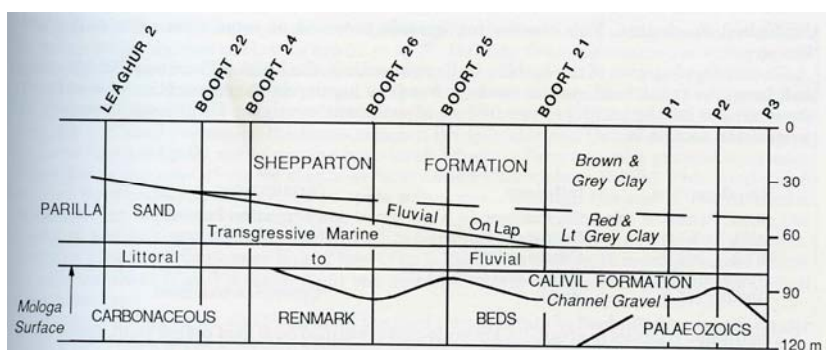


Figure 2.18 - Schematic diagram across the Pliocene shoreline at Yando

Dunn (1905) noted that the lead wash and gravel within the highlands was almost exclusively quartz, and recent drilling shows this to be the case on the plain. Of the many bores drilled on the plain, only one (Macorna 3) contained pebbles with lithologies other than quartz. This is not surprising, as it was drilled immediately to the north of the only known (buried) bedrock high rising from the Mologa Surface. The absence of lithic fragments other than quartz indicates a phase of intensive chemical weathering of the highlands prior to the deposition of the Calivil Formation. This weathering episode occurred at some time prior to the late Tertiary.

The confined section of the Loddon Valley passes into the Loddon Plain near Bridgewater, and, here, the trunk lead system emerges from the highlands. A cross-section at Woodstock about 20 km to the south, shows 103 m of sediment overlying Ordovician basement. A generalized section is:

Depth (m)	Lithology	Stratigraphy
0.00 – 49.4	Clay and sands	Shepparton Formation
49.4 – 87.2	Coarse sand	Calivil Formation
87.2 - 100.9	Coarse gravel	
100.9 - 103.0	Gravel and pebbles	
103.0 - 105.8	Slate	Ordovician basement

The infilled channel is about 1.5 km wide and contained 53.6 m of coarse sand, gravel, and pebbles. Further north at Bridgewater, a section shows the system as a 3.2 km wide valley, incised 61 m into Palaeozoic sediments, whose uppermost surface forms the Mologa Surface, which is itself about 60 m below the Loddon Plain. At this point, the valley has an asymmetric cross section, with an abrupt western boundary and a sloping eastern boundary. Infilling of the lead at Bridgewater shows a similar pattern to that found at Woodstock. Sand, gravel, and pebbles infill the valley trench to the level of the Mologa Surface, above which there is a marked change in character. Sand, silt, and clay occur to within about 40 m of the surface, above this level, the material is predominantly clay with sand lenses.

Beyond Bridgewater, the valley passes northward through the parishes of Yarraberb, Pompapiel, Calivil, Mologa, and Mincha West, towards Gunbower West in the Murray Valley. Throughout its length its character is that of a valley or trench incised into a pre-existing peneplain. At Bridgewater and Pompapiel, in the south, the valley bottom is formed by Palaeozoic sediments, but the system then passes onto, and is incised into, the carbonaceous Renmark Group. A section at Calivil shows lead gravel as occurring from 80-98 m and overlying 52 m of low permeability carbonaceous sand, silt, and clay of the Renmark Group. At this point the lead valley is cut 18 m into the Renmark Group, the top of which forms part of the Mologa Surface.

Northward, in the parish of Mologa, the Renmark Group is not present, and the valley is incised into Palaeozoic sediments (Figure 2.9). At this point the lead sequence is about 15.2 m thick and consists of coarse sand and gravel with minor clay seams. This situation continues into Gunbower West where a Murray Valley system is met and the Calivil Formation again overlies the Renmark Group.

2.5.5 Calivil Formation Beneath the Loddon Plain

Beyond Bridgewater, the coarse pebbly wash so characteristic of the Calivil Formation in the highland tract rapidly cuts out and is replaced by gravel. For instance at Calivil the Calivil 2 bore has the following lithology:

CALIVIL 2

Depth (m)	Lithology
0.0 - 51.8	Clay with minor shoestring sands
51.8 – 76.8	Dense sandy clay
76.8 – 95.1	Gravel with some minor white clay (Calivil Formation)
95.1 - 150.3	Ligneous sand and clay (Renmark Group)
150. +	Palaeozoic bedrock

Coarse sands and gravels comprise the basal unit in the lead valley along the entire length. At Bridgewater, the coarse clastics are 61 m thick, infilling the entire ancestral valley to the level of the Mologa Surface and spilling out over it. From Bridgewater, on passing northward, there is a gradual thinning of the gravel, so that at Calivil the sequence only averages 15 m - 18 m but remains fairly constant from that point northward (Figure 2.19). The downstream decrease in Calivil Formation thickness between Bridgewater and Calivil is seen as being in response to a gradually rising marine base level, which in turn caused a general upstream migration in the zone of coarse clastic sedimentation (Macumber, 1978b). This would also account for an upward fining seen in some sequences. Although the end of coarse clastic sedimentation is generally sharp,

some bores show an upward decrease in grain size in the sequence rather than a sharp change to finer-grained sediments, e.g., the Calivil bores, Macorna 2 bore, and Gunbower West 2 bore (Figure 2.19). Nevertheless, in most instances where this occurs, there is still a marked cut-out point for coarse clastic sediments of the Calivil Formation type.

Between Bridgewater and Pompapiel the lead trench is totally backfilled by coarse clastic sediments, and a sheet of sandy sediment has been deposited widely across the adjacent Mologa surface. This sand sheet plays an important hydrogeological role in that it provides an intermediate layer linking the main Calivil Formation aquifer with the shoestring sand aquifers of the overlying Shepparton Formation. As a consequence, any pressure buildup in the deeper Calivil Formation may be transmitted across the entire southern Loddon Plain, to be eventually dissipated at the surface by upward leakage through intersecting shoestring sands.

From the central Loddon Plain northward, the trench is not completely backfilled by coarse sediments; instead, the final backfilling is with a dense, red and grey mottled clay. This clay forms a virtual valley plug over the Calivil Formation gravels; it occurs in the Mincha West 10 bore, the Macorna 2 bore, and Macarthur's private bore (Mologa 8001) to the southwest of Pyramid Hill. Adjacent to the lead valley, bores such as Mincha West 11 and Loddon 2 show prominent red and grey mottling in the lower parts of their clay sequences, and this probably represents the same events as are recorded in the Yando area further west (see Chapter 2.7). Further downstream, the coarse sands and gravels deposited within the trench of the confined Loddon system merge with similar sediments, which are considered to have been deposited by an ancestral Murray system. In the Gunbower West 2 bore, drilled 8 km south of Cohuna, sand and gravel, occurring from 79 m to 103 m, are taken to represent deposits of a late Tertiary Murray River system, flowing from the east. As is the case in the central Loddon Valley, the gravels overlie the fine-grained, carbonaceous Renmark Group.

The change in fluvial province from the Loddon Valley to the Murray Valley is reflected in an abrupt change in water salinity in the Calivil Formation aquifer. In the Loddon Valley, salinity gradually increases downbasin to over 15,000 mg/l TDS at Mincha West; however upon entering the Murray Valley regime, the salinity of the Calivil Formation falls to 4,000 - 6,000 mg/l (Figure 4.6c, Table 4.3). Westward beyond Cohuna, the Calivil Formation is directly overlain by late Miocene to early Pliocene marine sediments - the Parilla Sand. This unit was originally defined by Firman (1965) in South Australia as a fluvio-lacustrine deposit; however, in Victoria it has been broadened by Lawrence (1975) to include the Victorian equivalents of the marine Loxton Sand of South Australia (Ludbrook, 1957). The Parilla Sand was deposited during the short-lived late Miocene to early Pliocene transgression which swept across northwestern Victoria.

Although the Gredgwin Ridge is the most easterly outcrop of the Parilla Sand in the Murray Basin, it has been traced eastward beyond Cohuna, under the Quaternary sediments of the Loddon Plain (Macumber, 1969). There is no marine sediment in the Wakool bores 36 078 and 36 102, situated in New South Wales to the north of Cohuna; the nearest marine sequences recorded in New South Wales bores are at Balranald, about 80 km to the west (Martin, 1977). It seems that the sea transgressed up the late Miocene Murray Valley, and at its height a deep marine embayment covered the lower Loddon Plain with its most easterly limits passing just inland of the junction of the Loddon and Murray river systems (Figure 2.20). At this stage the Murray Valley downstream from Cohuna, was drowned; the lower Loddon River system, upstream from this point, was estuarine. West of Cohuna, the Calivil Formation gravel is directly overlain by, and partly intertongues with the Kerang Sand Member of the Parilla Sand (Figure 2.21). In the Kerang area, the Kerang Sand is the basal transgressive unit in the marine sequence and consists of fine- to coarse-grained, micaceous sands formed in part by reworking of the Calivil Formation (see Chapter 2.9.1). The Kerang Sand is overlain by a clay and silt unit (Tragowel Member), and this in turn is overlain by a micaceous sandstone (Wandella Sandstone). This threefold subdivision of the Parilla Sand is discussed more fully in Chapter 2.9.1; it is taken to represent the form of the transgression-regression cycle at the extreme limits of the transgression in the Loddon Basin.

While the Parilla Sand outcropping on the Gredgwin Ridge is clearly marine, on passing eastward toward the limits of the transgression, deltaic and estuarine elements associated with Murray and Loddon river systems should also be present. No attempt is made to differentiate these various environments. The situation in the lower Loddon Valley is similar to that described by Taylor (1876) as occurring at Stawell, in western Victoria, where marine fossils were found overlying the auriferous lead sediments at the Welcome Rush, Poverty Hill and Four Post. Taylor commented that, before its subsidence below the sea, the original schist country on which the marine beds were subsequently deposited was a naturally undulating surface, with its own system of valleys and their tributary branches. They were then filled by the marine "drifts" which covered the older pre-existing fluvial "drifts". In summary, the gradually rising base levels that accompanied the late Tertiary transgression led to fluvial aggradation within the Murray and Loddon valleys. The nature of the aggradation is best seen in the Loddon Valley where the late Miocene Loddon River was confined to a deep trench incised 30 m to 60 m into the pre-existing middle Miocene Mologa peneplain. The trench was backfilled during the transgression and then buried under coastal plain deposits. In general, the trench sediments consist of a thick suite of fluvial gravel and sand. In the far north however, closest to the shoreline, the valley is not entirely backfilled with coarse-grained deposits, but, instead, the final material is a dense clay valley plug. On passing upvalley, the gravel sequence rapidly thickens to be at the highland front three times the thickness of the sequence when measured far out on the plain. An explanation of the downvalley wedging of the gravels is seen in the gradual rise in base level of aggradation, causing the deposition of finer-grained sediments in a gradually retreating flood plain environment which is developed as a coastal plain. Thus, as flood plain environments existed in areas marginal to the advancing sea, coarse-grained valley sedimentation continued uninterrupted upstream beyond the immediate influence of the rising sea level.

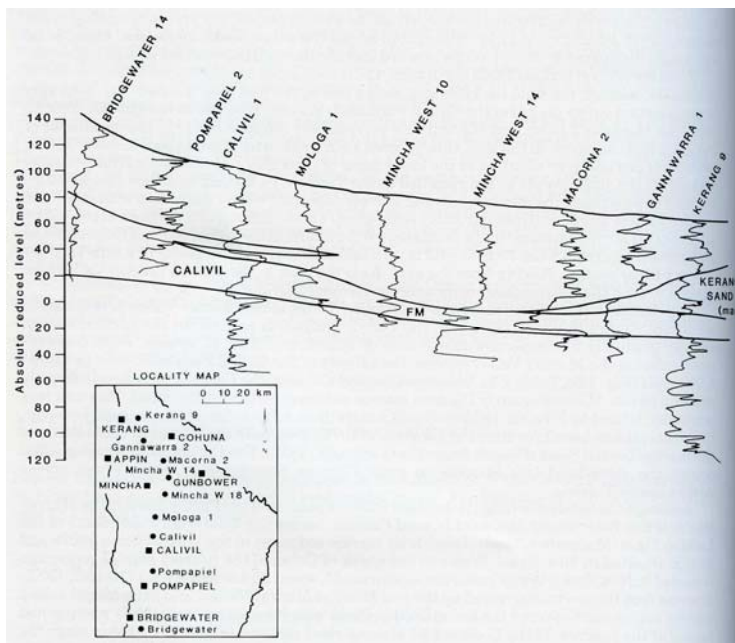


Figure 2.19 – Gamma log correlation of the Calivil Formation aquifer across the Loddon Plains

Figure 2.20 – Palaeogeography of the Loddon-Avoca plains in early Pliocene times (from Macumber, 1978b).

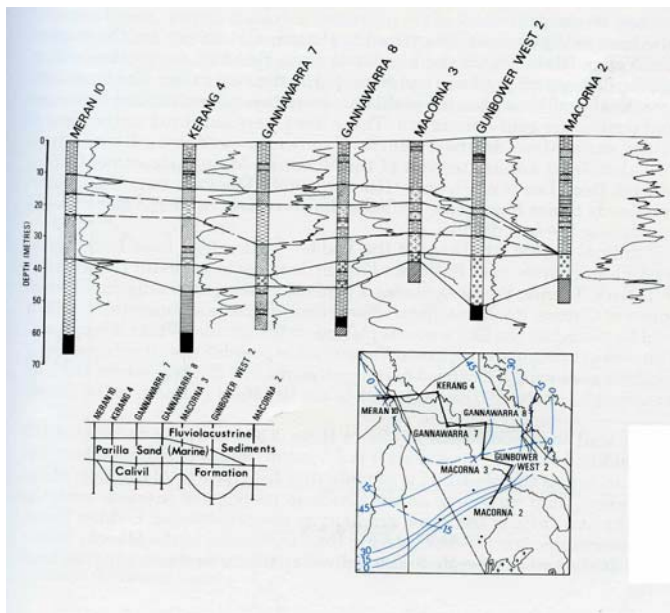
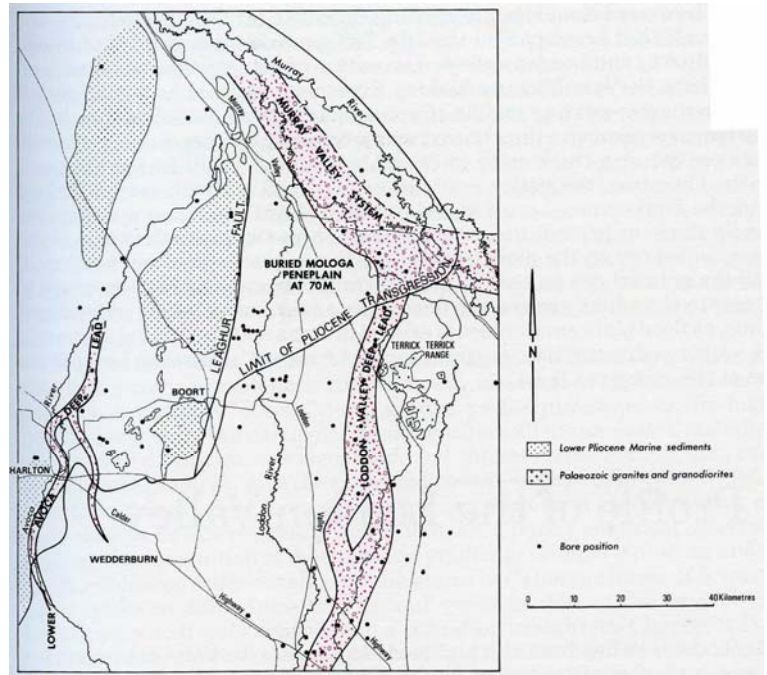


Figure 2.21 – Stratigraphic section through the lower Loddon Plain. Lower left is a generalized section, and lower right is a locality plan and isopach map of the Parilla Sand (from Macumber, 1978b).

2.6 Long Profile of the Loddon Valley Lead

The section along the Loddon Valley from the highlands to the late Tertiary sea provides a profile from which some estimation of the extent, if any, of late Tertiary uplift of the highlands relative to the plains may be gauged.

In Figure 2.22, the base of the Loddon Valley lead is plotted; also shown are the present surface of the Loddon Valley (from within the highlands and extending across the Loddon Plain) and the Mologa Surface, which predates and is incised by the lead valley. The basic data used to plot the longitudinal profiles within the highlands were taken from various bore lines drilled across the lead during the gold-mining era. These lines were surveyed at the time of drilling, and the position, surface level and basement level were measured for each bore. Much of the information is taken from annual reports of the Victorian Mines Department, from drilling records, and from Deep Leads of Victoria (Hunter, 1909). The data from the Loddon Plain north of the highlands comes from bores drilled either privately or by the Department of Minerals & Energy during the last decade.

A plot of the longitudinal profile (thalweg) of the Loddon Valley lead base from Midas (situated on the physiographic divide near Ballarat) 170 km downbasin to Mount Hope (at the northern end of the Terrick Terrick Ranges), shows a general concave upwards curve, with three major knick points at Clunes, Rodborough and Neereman. There is an uncertainty about what emphasis should be placed on the knick points plotted on the profile. While it may seem that they represent phases in uplift of the highlands, equally it is possible that their cause may be some lithologic controls producing a localized base level, as may be the case where the lead cuts a narrow gutter through the Palaeozoics adjacent to the Bradford granites in the parish of Neereman.

The overall fall of the lead base from Midas to Mount Hope is 311 m over a distance of 171 km, or a gradient of 1.8 m/km. In the highland tract it is 7.3 m/km (Midas to Bridgewater), but it is only about 0.63 m/km across the plains. It is notable that the thalweg of the base of the late Tertiary Loddon Valley, from its source on the divide to its point of junction with the Murray Valley system, has essentially the same gradient as the present-day Loddon River. Given the differing physiographic tracts when passing from the divide to the Murray River, there is, surprisingly, a fairly uniform depth of alluvial valley fill throughout, ranging from about 90 m to 120 m in thickness. This is in line with the observations of Wilkinson (1907), who noted that within the highlands the aggregate thickness of the alluvial sequence infilling the valleys was fairly constant. It is not only the case within the highlands but also out on the Loddon Plain, where it is unexpected, given the extremes of energies suggested by the widely different sediment loads of the late Tertiary lead and the more recent Quaternary Loddon River system. For instance, whilst the late Tertiary system transported gravel across the Loddon Plain, the Quaternary Loddon River plain was aggraded by fine-grained sediments and was little more than a basin of internal drainage. This is especially true northward beyond an east-west line from Calivil to Boort, where the depositional systems are either at or near base level, and with the exception of a few minor shoestring sands, carry almost entirely a suspended and soluble load. Similarly, while there is a slightly steeper gradient on the Mologa Surface between Bridgewater and Pompapiel, beyond this point the gradients of the Mologa Surface, the later Tertiary lead system, and the present-day Loddon Plain have essentially the same slope.

There is no obvious reason for the concordance of all these gradients; yet it seems unlikely that it is purely coincidental. However, any alternative explanation implies that this final gradient is virtually 'normal' for a graded profile plainward of the highlands, i.e., for the Loddon Plain there is a particular gradient where all the variables, including elevation, stream velocity and discharge, sediment load, and channel cross section, etc., capable of influencing the ultimate gradient, give the same end result - a slope of about 0.6 - 0.5 m/km. If this is the case, then this

gradient is an equilibrium gradient for a graded system on the Loddon Plain at, or close to, base level. While this is, in itself, a purely empirical observation of questionable validity, the recognition of these similar gradients is important to an understanding of events which influenced sedimentation on the plains in Cainozoic times. It is worth stressing that the only surface to differ from the 'normal' grade is that of the base of the Renmark Group sediments, which were deposited in a subsiding trough (see Chapter 2.3). The concordance of gradients of the three surfaces indicates that there has been no subsidence of the Loddon Plain since at least the time of incision of the lead system into the Mologa Surface in middle to late Miocene times, despite a marine incursion in the lower Pliocene. While supporting Wellman's contention that the late Cainozoic uplift of the Victorian highlands may only extend from the southern margin of the highlands to about 40 km north of the main divide (Wellman, 1974), this data goes further, suggesting that there has been no significant uplift of the Victorian western highlands since before the development of the Mologa Surface in middle Miocene times.

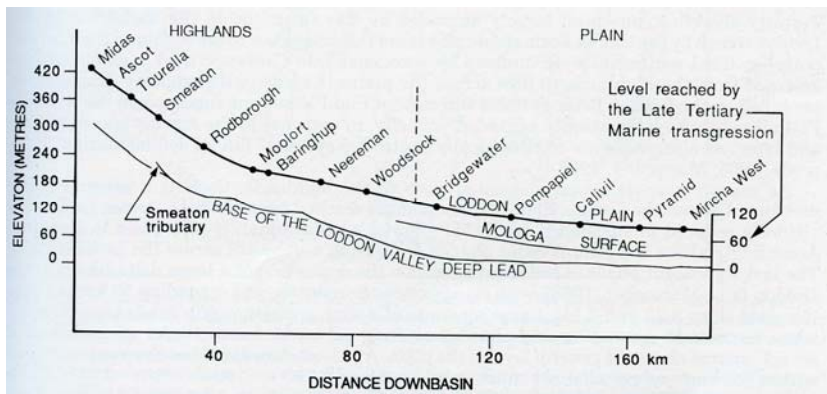


Figure 2.22 - Thalwegs (long profiles) of the base of the Loddon Valley lead, the Mologa Surface and the present surface of the Loddon Valley

2.7 Shepparton Formation

The Shepparton Formation was defined by Lawrence (1966) as being a thick sequence of fluvial and lacustrine sediments overlying the Calivil Formation. The sediments are principally brown and grey mottled clay and silt, with thin beds of fluvial sand - the shoestring sands of the prior streams. The Shepparton Formation is distinguished from a younger Coonambidgal Formation (Lawrence, 1966) by the latter being essentially, deposits of the existing streams or their recent ancestors in the Murray Basin. On the Loddon Plain, a clear distinction can often be made between a lower sequence of red and light grey clays and an overlying suite of brown and grey clays, the boundary between the units being at times very sharp, but becoming less distinct upbasin towards the highlands.

Even so, the distinction was clear in the buried Powlett Plains valley, where a feature of the Shepparton Formation is its subdivision into two units: a lower unit of red-brown and light grey mottled clay with lateritic grit, and an upper unit of yellow-brown and grey clay. This distinction was also noted in sections further north at Yando and Yarrowalla, where very distinct, red-brown and light grey clays occupy the same stratigraphic interval as the Parilla Sand immediately to the north - where borehole data show them as also onlapping the Parilla Sand (Figure 2.18). The reddish, mottled clay is interpreted as being a nonmarine, lateral facies equivalent to the Parilla Sand; it was probably deposited in a coastal plain environment. As was shown earlier (see Chapter 2.5.5), the same reddish, mottled clay overlies the Calivil Formation in the Loddon Valley trench, north of Mologa, where it forms a final valley-fill deposit and then spills over onto the surrounding Mologa Surface. The red and grey mottled clay is an easily distinguished lithological unit over many parts of the Loddon Plain and is therefore regarded as a discrete member of the Shepparton Formation - the Yando Clay. The colour distinction between the red clay and yellow-brown clay is seen as having climatic and environmental implications, with the reddish colours suggesting that conditions existed at the time which were conducive to ferruginization. The change to yellow-brown colours is interpreted as marking the termination of these conditions. A tentative correlation may be made between the ferruginizing events on the Loddon Plain and the conditions which led to the formation of the ferruginized and silicified Karoonda Surface, developed on the Parilla Sand after its exposure to subaerial weathering (Firman, 1966). However, whereas the latter event is generally seen as essentially a postregressional feature, the evidence from the nonmarine sequences on the Loddon Plain indicates that these same conditions were present during the time of the transgression. While the central lower Loddon Plain, landwards from the late Tertiary shoreline, has been largely aggraded by clay (discounting the backfilling of the Loddon trench by the Calivil Formation), this is not the case closer to the highland front, where coarse-grained sediments were dumped by successive late Cainozoic river systems as they emerged from the highlands to flow across the plains. Extensive distributary development prevailed on the Loddon River systems throughout Plio-Pleistocene times, when the Riverine Plain was being systematically aggraded: initially, in response to the marine transgression, and later, as a consequence of climatically controlled cycles of fluvial sedimentation (Lawrence, 1966; Macumber, 1968).

As coarser materials were deposited close to the highlands, there is a gradual fining northward across the Loddon Plain, until a point is reached beyond which stream loads were virtually reduced to suspended load; this allowed a predominantly clay plain to build up. Accordingly, there is a lateral facies change on passing northward across the Loddon Plain. The last significant phase of sedimentation saw the deposition of a large delta-like fan (the Loddon fan - Macumber, 1968), with its apex near Serpentine and extending 30 km out onto the plain (Figs 2.23 and 3.11). Large amounts of coarse-grained quartz sands were dumped when large-scale crevassing and channel shifting occurred. Levee banks were deposited several metres above the general level of the plain. A radiocarbon date from charcoal contained within the more recent channel sands gave an age of 8,790 $\pm$ 120 BP (SUA - 1320), which coincides with a phase of high lake levels at Kow Swamp on the lower Loddon Plain (Macumber, 1977).

One important consequence of the distribution of Shepparton Formation lithologies across Loddon Plain is that the Calivil Formation, as an aquifer, is semiconfined to semiunconfined in the southern Loddon Plain (where shoestring sands are common) but becomes increasingly confined toward the north, where the Shepparton Formation is finer-grained once the late Tertiary shoreline is crossed the Calivil Formation is overlain by another aquifer, the Parilla Sand, which provides no real impediment to upwards flow; again, the aquifer is, at most, semiconfined.

2.8 Bookpurnong Beds

The late Tertiary transgression commenced with the deposition of a thin sequence of fossiliferous glauconitic marl, referred to by Ludbrook (1957) as the Bookpurnong Beds. In Victoria, the Bookpurnong Beds are fossiliferous clay, clayey marl, calcareous silt and clayey sand, which are commonly glauconitic. They have a much broader lithology than beds described by Ludbrook from the South Australian type sections along the Murray River. They are well documented by Gloe (1947) and Lawrence (1966) in the Murrayville region of far northwestern Victoria, where they were shown to thicken on the downthrown side of the Danyo Fault. The Bookpurnong Beds unconformably overlie the Duddo Limestone and the Winnambool Formation in the western parts of the Murray Basin but tend to be missing further east and were not identified in the Tyrrell Basin (Chapter 2.9.2).

2.9 Parilla Sand

Covering the whole of the marine Murray Basin is a blanket of sand and sandstone, having an average thickness of about 60 m. In Victoria, the upper surface of the sandstone forms long prominent ridges trending north-south, which are generally regarded as having been formed as strandlines along a retreating late Tertiary sea (Blackburn, 1962).

Ludbrook (1961) described sands from a type section at Loxton in South Australia, where they are "characteristically bright yellow cross-bedded micaceous sands, grits and silty sands...deposited in estuarine or shallow water at the end of the Miocene". This sequence is 15.2 m thick and contains molluscan faunas, including *Donax depressa*, *Glycymeris cainozoica* and *Nucularia sp.*

The Loddon fan commences near Serpentine after the ancestral Loddon River leaves its previously incised meander tract and spreads as a distributary system across the plain. The head of the fan approximately coincides with the hinge line between regional groundwater recharge and discharge zones.

Present day groundwater discharge areas and fossil discharge areas are indicated by the large number of lakes and lunettes in the northern areas, while the nearby Barr Creek is also a groundwater discharge feature (Chapter 4)

Firman (1966) split the Loxton Sands into two units - the Lower Beds (estuarine and equivalent to the sand of the type section at Loxton) and Upper Beds (fluviolacustrine), with a disconformity between. Furthermore, he recognized another sand unit the Parilla Sand, disconformably overlying the Loxton Sands and having similar lithology (though finer-grained) but, unlike the latter unit, having little or no cross-bedding. All three sand units together were about 61 m thick. The sand sheet occurred widely in Victoria - as far east as Nyah. The Parilla Sand unit was considered by Firman to be fluviolacustrine and the complete sequence as being Pliocene.

Shortly afterwards, Lawrence (1966) defined the Diapur Sandstone, which he considered to be a facies variant of the Loxton Sands - the junction lying "somewhere in the South Australian portion of the Murray Basin". The Diapur Sandstone differed from the Loxton Sands in structure (it has no strong cross-bedding), lithology (it is medium- to fine-grained and only lightly micaceous), and lack of fossils. Although he found it unfossiliferous, Lawrence suggested that the lower part of the Diapur Sandstone might have been deposited under offshore to littoral conditions; and that the upper material forming ridges was a shoreline deposit. However, the Diapur Sandstone, in description and areal extent almost exactly coincided with the Parilla Sand; and, on realizing this, Lawrence adopted Firman's prior nomenclature. The importance of this unit, firstly, as the downbasin marine equivalent of the fluvial Calivil Formation (for the purposes of regional groundwater flow) and, secondly, because it forms the regional unconfined aquifer throughout northwestern Victoria, made it imperative that emphasis should be paid to it in this study of groundwater flow. In the early parts of the study, the nature of the sand sheet at its eastern (inland) limits was investigated, and its stratigraphic relationship to fluviolacustrine sequences of the Riverine Plain was established by a comprehensive drilling program.

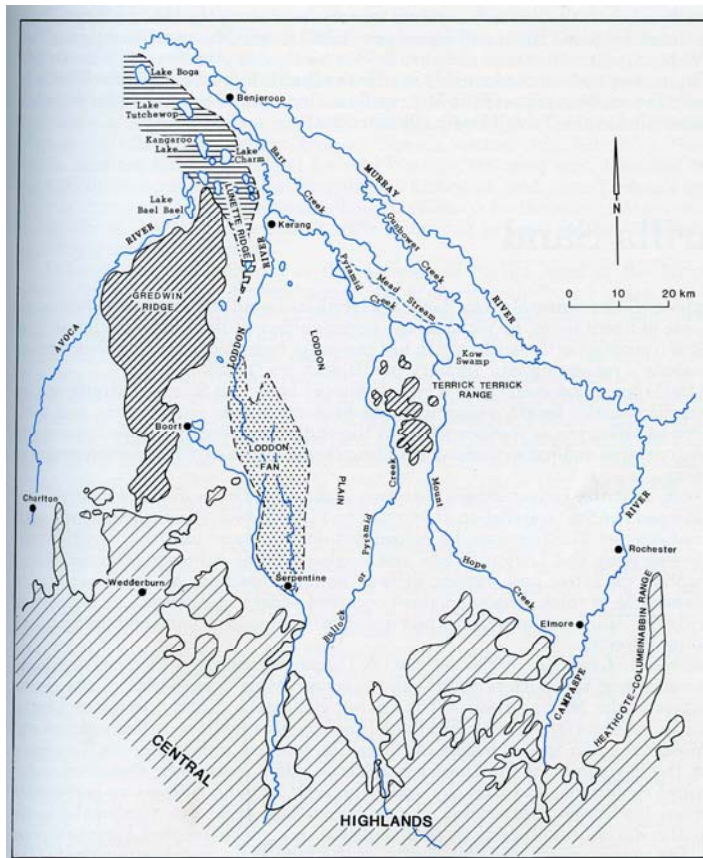


Figure 2.23 - Geomorphology of the Loddon Plain (from Macumber, 1968).

All earlier descriptions of the Loxton-Parilla-Diapur Sand/Stone were from the western of the Murray Basin, although the latter two units were extended to the east by both Firman (1966) and Lawrence (1966). Working at the eastern limits of the sand/stone sheet on Gredgwin Ridge, Macumber (1969) found marine fossils both at the base of the sequence and at the top of the sequence (Macumber, 1978b), indicating that at its inland limits the consequence was entirely marine. Fauna from the top of the sequence (Figure 2.24a) contained the 'noid *Lovenia woodsi*, which elsewhere in Victoria is considered to be Upper Miocene to lower Pliocene. Quarries on the Gredgwin Ridge were commonly fossiliferous; in many instances the bedding planes were covered with disarticulated pelecypod valves, suggesting beach or nearshore environment. Molluscs present included *Dosinia* (sensu lato), *Garia* and *Ilina* sp. Various leaves associated with the marine fauna were also found in material from the Gredgwin Ridge quarries. The data from the Gredgwin Ridge is in agreement with the work Taylor (1876), who investigated a number of fossiliferous localities in sands onlapping the highlands at Stawell. Marine molluscs, described by Taylor, included *Lepralia stawellensis*, *rupa wormbetiensis*, *Tellina krausei*, *Nucula martae* and *Turritella*. Since the Loxton ds were not known to extend far into Victoria, the sand unit described by Firman was essentially the same unit as that described by Lawrence, and, since Firman's terminology had priority, the name Parilla Sand was adopted in Victoria for the entire sand/stone sheet. For the remainder of this study the term Parilla Sand will be used in the Victorian sense unless specifically stated otherwise.

Although the vast majority of the outcrops of Parilla Sand in the Murray Basin are essentially flat-lying fine-grained sandstone as described by both Firman and Lawrence, the sequence found on the Gredgwin Ridge is more complex than either Firman's description of the Parilla Sand or Lawrence's Diapur Sandstone. This is to be expected, since the area forms the transitional zone spanning the marine/nonmarine boundary of the basin and units show a diversity not apparent in

the areas further west. The complete clastic unit of over 60 m has elements which are unlike either the defined Parilla Sand or Diapur Sandstone in that it is, places, strongly cross-bedded (Figure 2.24b) and consists of fine- to coarse-grained, strongly micaceous sandstone and sandy grit (Macumber, 1969). Upper parts of the sequence, in fact, better fit the description of the Loxton Sands, originally defined by Ludbrook as "cross-bedded, gritty, micaceous sands, with shelly beds". It seems likely that the Loxton Sands in South Australia and units within the Parilla Sand on the Gredgwin Ridge are a distinct lithofacies variant which is found on the basin margins. A further feature of the Parilla Sand on the Gredgwin Ridge is the presence of heavy minerals bands up to 1 m thick. The minerals 'present an ultrastable suite, consisting of zircon (10-30%), ilmenite (40-65%), rutile (5-11%), leucoxene (2-7%), and smaller amounts of tourmaline, monazite, chromite, and other stable minerals (all percentages being in terms of total heavy mineral content). These are beach lacer or shallow marine placer deposits (Macumber, 1969) and are interbedded with fossiliferous sandstone. The monazite makes the bands mildly radioactive, and this often shows as peaks on gamma logs which pass through the Parilla Sand.

Apart from the section line at the lower end of the Loddon Valley (Figure 2.21), two further sections were drilled across the late Tertiary shoreline to determine the relationship between the marine Parilla Sand and the nonmarine coastal plain sequences. The sections - an east-west section at Mincha West (Figure 2.6) and a north-south section at Leaghur (Figure 2.18) - show that the Parilla Sand was deposited across the Mologa Surface and wedged out in the central Loddon Plain. The nonmarine equivalents are commonly red and grey mottled clays - the Yando Clay, which onlaps the marine sequences in both sections. Similar sediments overlie the Calivil Formation in the Loddon lead trench to the east but are replaced by the Parilla Sand further downbasin.

As sea level rose, there was an initial backfilling of the Loddon Valley trench, followed by a more widespread alluviation of the Loddon coastal plain. The level of the top of the Parilla Sand sequence in the Loddon 1 bore is 61 m above the present-day sea level (and); and this gives an approximate sea level reached at the peak of the transgression (bearing in mind that there is no evidence of subsidence of the Loddon Plain arising from the transgression nor was there uplift following it). It seems that the shoreline remained fixed for some considerable time at its eastern limits, where a fine balance was reached between sea level rise and fluvial aggradation of the coastal plain. The sea retreated from the Murray Basin in the early Pliocene, leaving the Riverine Plain in Victoria as essentially a basin of internal drainage.



Figure 2.24a – Parilla Sand exposure in a quarry, 13 km west of Kerang, where marine mollusks and the echinoid *Loveia woodsii* were obtained. A thin, darker band in the middle of the sequence is a heavy mineral layer.



Figure 2.24b – Uppermost unit of the Parilla Sand exposure showing cross-bedded sands and sandy grits. Heavy mineral bands (not shown) occur throughout the sequence, which forms the Wandella Sandstone Member of the Parilla Sand.

2.9.1 Subdivision of the Parilla Sand

At its inland limits, beneath the fluviolacustrine sediments of the Loddon Plain, the Parilla Sand may be subdivided on lithologies into three members: the uppermost, Wandella Sandstone; intermediate, Tragowel Member; and lower, Kerang Sand Member. The Kerang Sand and the Wandella Sandstone are seen as transgressive and regressive sand sheets, and the Tragowel Member is seen as a slightly deeper water deposit.

Kerang Sand. In the Loddon Valley, the Bookpurnong Beds are absent from the sequence, and the Kerang Sand (Macumber, 1978b) is the basal transgressive unit, directly overlying the fluvialite Calivil Formation, or in its absence the Renmark Group, or at times sitting directly on weathered Palaeozoic sediments, as is the case in the north central Loddon Valley. The Kerang Sand does not outcrop, and the type section is taken from the Meran 10 bore near Kerang (Figures 2.21 and 2.25), where it occurs from 47 m to 74 m and there overlies the Renmark Group. The Kerang Sand consists of up to 30 m of medium- to coarse-grained quartz sand, partially derived from reworking of the underlying Calivil Formation. Where it directly overlies the latter sands in the Cohuna-Kerang area of the lower Loddon Plain, it is often difficult to draw the boundary between the Calivil Formation and the Parilla Sand. The palaeogeographic interpretation of the lower Loddon Valley in early Pliocene times has the Pliocene Loddon River junctioning with a Murray Valley system and flowing into the Tertiary sea near Cohuna (Figure 2.20). The marine Kerang Sand passes upbasin into the upper portion of the Calivil Formation, the boundary between the two sand units being probably diachronous.

Tragowel Member. Beneath the Loddon Plain, the Kerang Sand is underlain by a sequence of white clay and silt - the Tragowel Member (Macumber, 1978b) of the Parilla Sand. The unit is up to 25 m thick and was not observed in outcrop in Victoria, although similar white clays were observed overlying units of the Loxton Sands along the Murray River trench in South Australia (Figure 2.26). The type section for this member is the Meran 10 bore on the shore of Lake Wandella in the parish of Meran, where it occurs at a depth of 33 m to 47 m. The abridged lithology in Meran 10, situated about 2 km west of Kerang, is as follows:

MERAN 10

Depth (m)	Lithology
0 - 20	Mottled fluviolacustrine clay
20 - 33	Fine-grained micaceous sandstone, slightly ferruginous, with small amounts of ilmenite, rutile and zircon (Wandella Sandstone)
33 - 47	White clay (Tragowel Member)
47 - 74	Medium- to coarse-grained, micaceous sand (Kerang Sand)
74 - 212	Carbonaceous silt, sand, clay and minor lignite(Renmark Group)

The threefold subdivision of the Parilla Sand is easily recognized on the gamma logs. The example given below (Figure 2.25), from Meran 10, also shows a 'kick', at 24 m; this is interpreted as a zone of heavy mineral concentration; a second 'kick', at 74m, is interpreted as a soil developed on the top of the underlying carbonaceous unit. In the Loddon Valley, the Tragowel Member of the Parilla Sand varies from a fine-grained, light grey to white quartz silt, to a silty clay/stone, in which clean transparent silt-sized quartz grains are contained in a clay matrix. On passing westward under the Gredgwin Ridge, the clays pass laterally into micaceous silty sands and sandy silts, with occasional clay bands. They are similar in texture to the finer parts of the overlying Wandella Sandstone but are more uniform, flat-lying, and lack the cross-bedding of the Wandella Sandstone. In Mawson's Quarry to the west of Kerang, they are fossiliferous, containing essentially the same fauna as in the overlying Wandella Sandstone.

Wandella Sandstone. This is a fine- to coarse-grained, micaceous sandstone (Macumber, 1978b), with minor gravel; it has occasional bands of heavy minerals. It outcrops on the Gredgwin Ridge, where it is fossiliferous and in places shows moderate cross-bedding (Figure 2.24b). It is, in general, sparingly fossiliferous, although some individual beds are richly fossiliferous. Westward of the Loddon Valley, lithologies within the Parilla Sand are less uniform and, therefore, less easily correlated. Variations include sequences consisting of sands and sandstone throughout; sand/stone sequences overlying silts and fine sands; and sequences similar to that in the Loddon Valley, where a white clay unit is sandwiched between two sand bodies. In two closely spaced lines of bores, drilled at Quambatook in the Avoca Valley and at Wycheproof, sequences similar to that of the Loddon Valley were found. A white clay, although not as well developed as in the Loddon Valley, formed an intermediate unit between two sand units. Although not directly correlatable with the Tragowel Member of the Loddon Plain, it is, nevertheless, seen as a western equivalent of this member. A feature of the Parilla Sand is the presence of large amounts of iron oxides throughout the sequences, probably arising from the remobilization of lateritic ironstones, followed by their redeposition in the upper recharge parts of the basin by through-flowing groundwater.

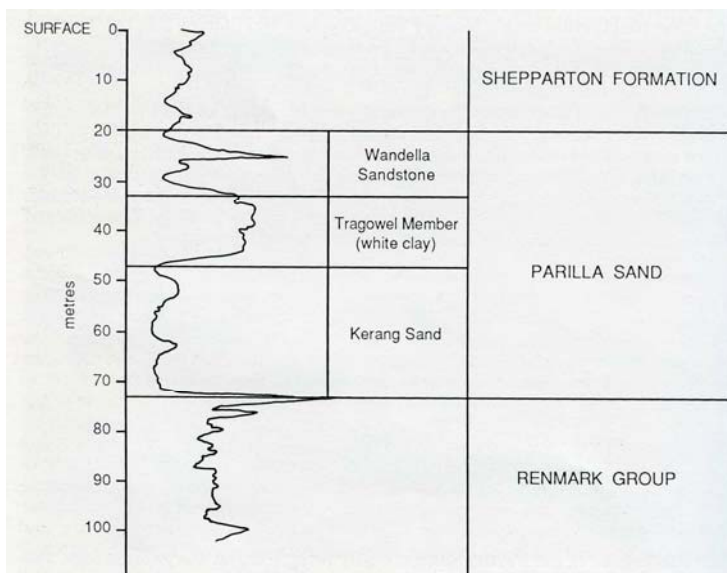


Figure 2.25 – Meran 10 gamma log, showing the threefold division of the Parilla Sand under the Loddon Plain

2.9.2 The Tyrrell Basin

Borehole sections in the Tyrrell district of the central Mallee show that the area covered by the three lakes, Tyrrell, Wahpool and Timboram (Figures 2.28, 2.29 and 2.30), is a broad, shallow downwarp in the Tertiary sediments lying immediately to the east of the Tyrrell Fault. The basin formed by this downwarp is about 35 km in diameter. The vertical movement of the warp is 30 m, seen by the displacement of the base of the Parilla Sand in bores to the west of Lake Tyrrell. A similar displacement can also be seen in the boundary between the upper and lower units of the Parilla Sand. On passing southward and eastward from Lake Tyrrell, the base of the Parilla Sand rises again, with the edge of the basin probably coinciding with a line of depressions a few kilometres to the east of Lake Timboram. The areal extent of the depression is readily seen by the distribution of the Blanchetown Clay, a lacustrine sequence of clay, silt and fine sand, which partially infilled the basin, and probably was a major factor in its subsidence (Figure 2.32). The basin was recognized by Hills (1939) as a distinct physiographic entity, which he called the Tyrrell Basin.

2.9.3 Lithology of the Parilla Sand in the Tyrrell Basin

In the Tyrrell Basin, the Parilla Sand may be subdivided on lithological grounds into two units - an upper coarse-grained sequence, consisting of fine- to coarse-grained sands with occasional gravels and a lower sequence of medium- to fine-grained sands and silts and interbedded, thin clay layers. The division between the two units is readily discerned from the gamma logs and occurs throughout the Tyrrell Basin. A further distinction is that the sequence has strongly oxidized colours towards the top but passes rapidly into a dark grey to black lower zone, which may at times be carbonaceous and pyritic, as in the Bourka 1 bore. The colour zonation does not coincide with the boundary between coarser and finer parts of the Parilla Sand. The colour is seen as being largely due to the presence of finely disseminated sedimentary organic matter, and, at times, wood particles associated with pyrite are present in the sequence, as was the case in the Bourka 1 bore. It is likely that iron sulphides also contribute to the dark colouration of the sediments. The darker colours generally, but not always, coincide with reducing conditions in the aquifer, that is, where pH values are neutral and Eh values are low or negative. The redox zonation in the Parilla Sand and its relationship to colour zonation is discussed in detail in Chapter 7.7 and following.

The presence of significant amounts of organic matter in the Parilla Sand is unusual, although thin coal seams a few centimetres thick have been occasionally observed in bores elsewhere throughout the Murray Basin, for instance, in the Koorangie 15 and 16 bores on the Gredgwin Ridge, in the Boinka 10002 bore at Boinka, and in the Goonegul 10001 and Mournpoul 2 bores in the Raak boinka. The sequences drilled through the Parilla Sand closer to the basin margins show essentially oxidized colours throughout; however, nearer to the Tyrrell Basin, the Kooem 1 bore, drilled on the Cannie Ridge at Kooem, about 60 km east of Lake Tyrrell, showed reduced colours from about 24 m below the surface to the base of the Parilla Sand, a further 67 m down. A similar change was observed in the Lalbert 1 bore, 60 km to the southeast of Lake Tyrrell, although no comparable carbonaceous sequence was found to equate with the sequence from the Tyrrell Basin. The presence of oxidizing conditions throughout the Parilla Sand sequences in areas closer to the basin margins (while reducing conditions are preserved nearer the centre of the basin) is seen as reflecting stages or zones within the regional groundwater flow system, with recharge taking place upbasin and discharge taking place towards the basin centre. The recharging oxygenated waters produce oxidizing conditions in the recharge areas, but the effects of this fade on passing downbasin (cf. Edmunds, 1973). The carbonaceous sands from the Tyrrell Basin and, to a lesser extent, the thin coal seams found elsewhere in the Parilla Sand are interpreted as indicating deposition in a paralic environment rather than in a strictly marine environment.

Therefore, while the Parilla Sand from the eastern Mallee is largely seen as a shallow marine to beach deposit, there is in the Tyrrell Basin evidence for the existence of paralic elements occurring deeper within the sequence.



Figure 2.26 – White clays overlying the Loxton Sands in a Murray River cliff section near Loxton.

These clays are tentatively correlated with the Tragowel Member of the Parilla Sand, as found in borehole sequences on the Loddon Plain

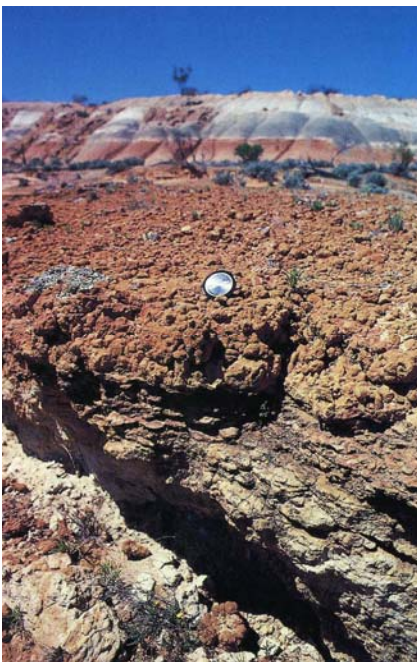


Figure 2.27 – The ferruginized and silicified top of the Parilla Sand, as exposed at Chowilla in S.A. The overlying sediments in the background are the Blanchetown Clay.

Not the different forms of ferruginization between that of the Karoonda Surface and that shown in Figure 2.26. While the former is seen as being of pedogenic origin, the iron oxide bands in Figure 2.26 are interpreted as being of groundwater origin.

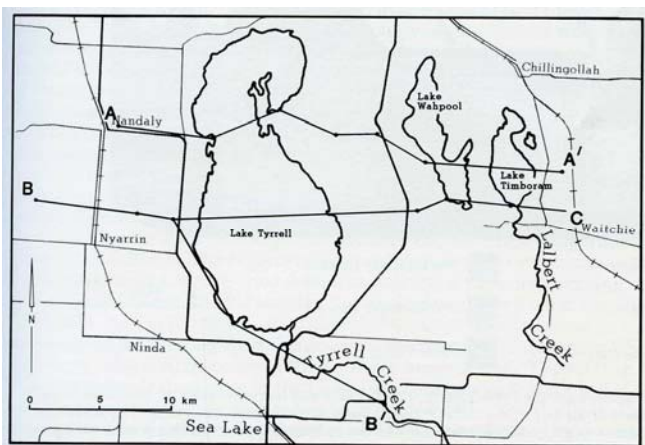


Figure 2.28a - Geological section through the Tyrrell Basin

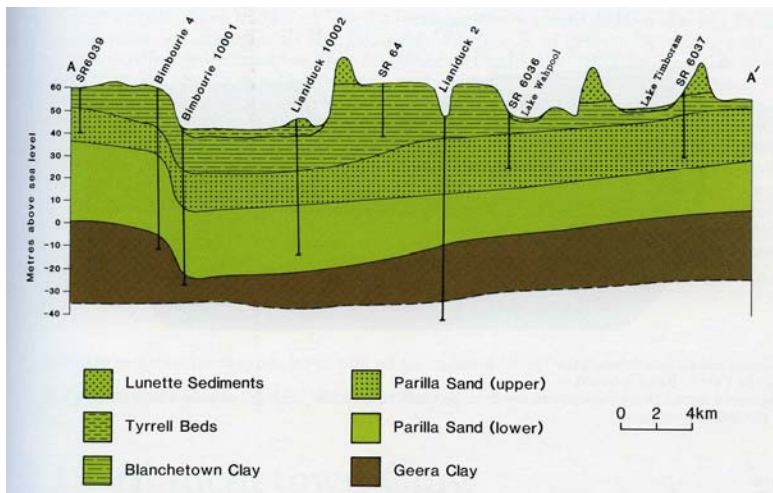


Figure 2.28b - Positions of section lines through the Tyrrell Basin

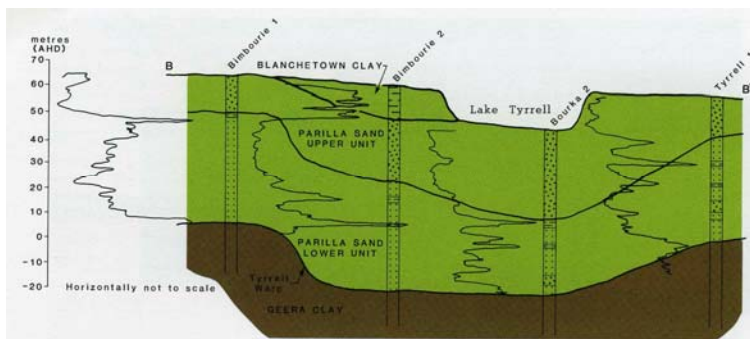


Figure 2.29 - Geological section through the Tyrrell Basin, giving the lithological and gamma log data on which the subdivision of the Parilla Sand is based.

The stratigraphic correlations shown here are from borehole to borehole, with the corresponding gamma logs appearing to the left of each bore.

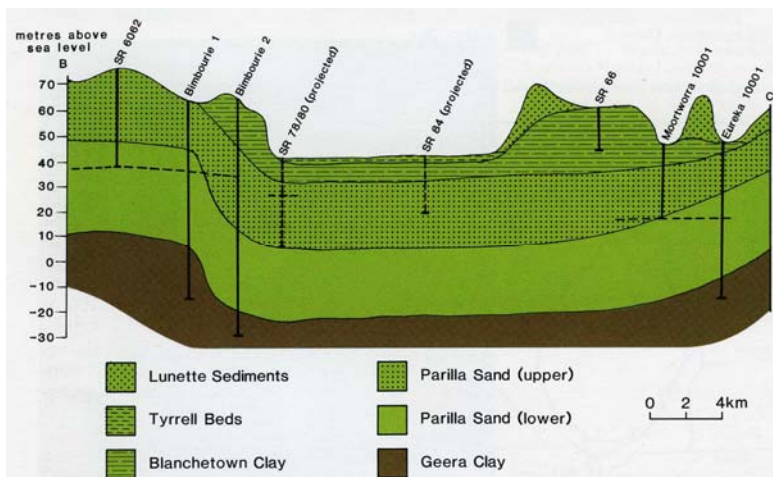


Figure 2.30 - Stratigraphic section through the Tyrrell Basin.

The Tyrrell Fault (lying between the Bimbourie 1 and 2 bores) forms the western limits of the basin and controls the position of the Blanchetown Clay. The horizontal broken line in the section separates brightly coloured oxidized sediments from sediments having grey to dark *grey* reduced colours.

2.9.4 Ferruginization of the Parilla Sand

The uppermost surface of the Parilla Sand is varyingly silicified and/or ferruginized (Firman, 1973; Lawrence, 1966) and has been called the Karoonda Surface by Firman (1973). Where it is ferruginous, the surface may be covered with small limonite pebbles and pisolites, as at the outcrop of Parilla Sand known as The Caverns, in the western part of the Big Desert. The ferruginization associated with the Karoonda Surface is of pedogenic origin, and the soil eloped on the Parilla Sand has been described by Coaldrake (1951) as a lateritic podsol. this pedogenic event should not be confused with a second form of ferruginization, which is widespread in the Mallee, and is commonly well developed on outcropping Parilla Sand where occurs low in the landscape. Here, the sands are cemented by iron oxides deposited from groundwaters (Figure 2.26). This process is discussed in Chapter 7. The two forms of ferruginization can be seen in section at the Chowilla site (South Australia), where the Parilla Sand forms cliffs to the Murray River. At the top of the Parilla Sand is the Karoonda Surface, littered with iron oxide pisolites (Figure 2.27), while at the base of the cliff close to river level the Parilla sand is cemented by iron oxide deposited from groundwaters. Ferruginization by groundwater an ongoing process; for example, at the Raak boinka, sands overlying the Blanchetown Clay have zones of heavy ferruginization several metres thick (Figure 7.20).

2.10 Blanchetown Clay

Plio-Pleistocene, lacustrine deposits occur over a wide area of the Murray Basin. Brown (1945) recognized the existence of a large lake occupying the central west Murray Basin. Deposited within the lake was a sequence of olive-grey clays and silts, with minor interbedded fine sands, which Firman (1965) called the Blanchetown Clay. He referred to the lake as Lake Bungunnia. In northwest Victoria, the Blanchetown Clay was shown by Lawrence (1966) to infill depressions or corridors between north-south ridges in the marine Parilla Sand, suggesting a series of lakes rather than a single large lake. These may have coalesced into one huge lake with islands of Parilla Sand, when water was at a higher level. Where it has been mapped in detail at its southern margins, the limits of the Blanchetown Clay are seen to be strongly influenced by basin structures, such as the Pinnaroo Block in South Australia (Firman, 1965), and the Sunset Strip and Tyrrell Fault in Victoria (Figures 2.31 and 6.2). In the latter instance, the thick Blanchetown Clay sequence has been deposited contemporaneously with the sagging of the Tyrrell Basin. The thickest measured sequence (34 m) occurs in the Mournpoul 2 bore in the Raak boinka, where it too is seen as indicating basement sagging in a manner analogous to the Tyrrell Basin. In this section, lignite occurs at the base of the Blanchetown Clay.

Within the Tyrrell Basin, the Blanchetown Clay was deposited in a single large lake, the perimeter of which roughly coincides with the margins of the basin (Figure 2.30). The thickest sequences (about 20 m) occur in the northern areas of Lake Tyrrell (Figure 2.32), with a gradual thinning towards the south. In the far south of the lake and in the bed of the Tyrrell Creek, the thin Blanchetown Clay sequence has been eroded away, and the floor of the lake is cut directly into the Parilla Sand. An Zhi-sheng et al. (1986) established, by means of magnetic reversal stratigraphy, a chronologic framework for the Blanchetown Clay. Using two stratigraphic sequences - from Chowilla in South Australia and from Lake Tyrrell - the Blanchetown Clay was shown to range in age from below the Matuyama - Gauss boundary, at about 2.4 m.y. (Upper Pliocene), up to the Bruhnes normal epoch, which commenced about 700,000 years ago. At Lake Tyrrell, the boundary between the magnetically reversed Matuyama epoch and the magnetically normal Bruhnes epoch coincides with the boundary between the Blanchetown Clay and an overlying 4 m to 5 m thick sequence of gypseous clays - the Tyrrell Beds (Teller, Bowler and Macumber, 1982). The incoming of the Tyrrell Beds marks the onset of saline conditions at Lake Tyrrell.

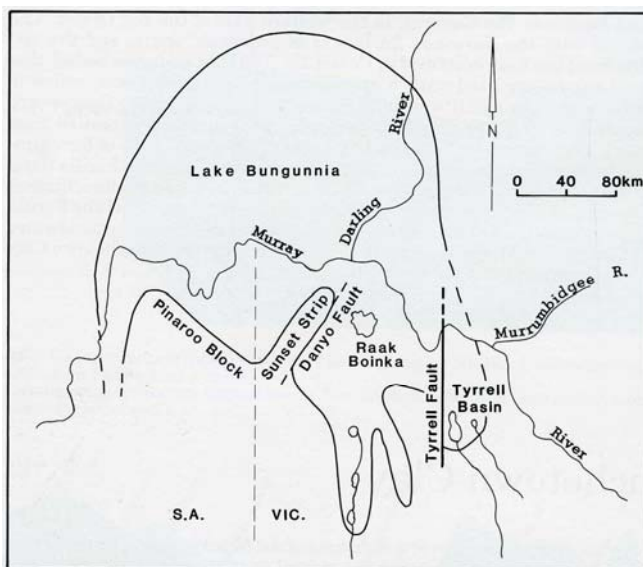
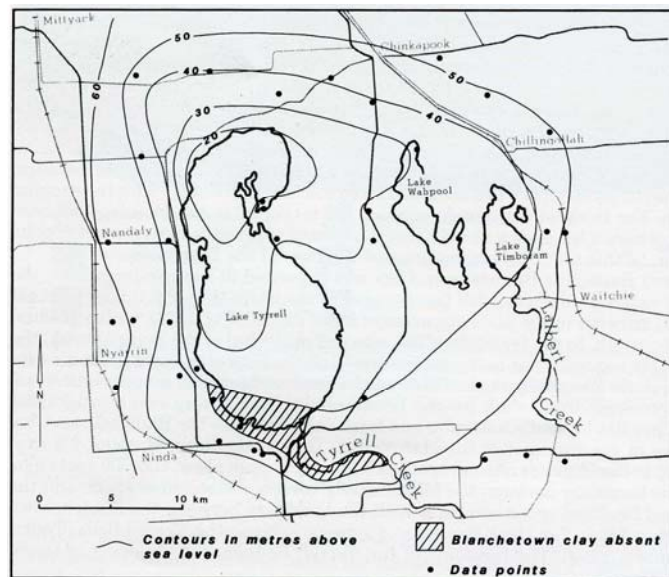


Figure 2.31 – Extent of the Plio-Pleistocene Lake Bungunnia (after Firman, 1965; Lawrence, 1975; Bowler, 1980; and Macumber, 1983a).

Figure 2.32 – Structure contours of the base of the Blanchetown Clay – Tyrrell Basin, northwestern Victoria



2.11 Middle-Late Pleistocene Aeolian Activity

It is notable that saline conditions, conducive to production of the Tyrrell Beds lithofacies, are a necessary requirement for lunette generation, and indeed an extensive system of varying aged lunettes overlies the Blanchetown Clay to the east of Lake Tyrrell. However, lunettes are just one of a number of aeolian features in northwestern Victoria and southeastern South Australia which date to this period; for instance, the aeolian Woorinen Formation (Lawrence, 1966), with its six palaeosols (Churchward, 1963), and elements of the Lowan Sand (Lawrence, 1966) both overlie the Blanchetown Clay. Further to the west, the regressive dune sequence, the Bridgewater Formation (Boutakoff, 1963), found between Narracorte and Robe, with the exception of the most inland East Narracorte Range, has everywhere a normal Bruhnes magnetic stratigraphy (Cook et al., 1977).

Both inland and at the coast, conditions were clearly conducive to extensive aeolian activity during the period following the demise of Lake Bungunnia. Bowler and Magee (1978) identified pelletal clay structures in the Woorinen Formation which indicate that these sediments were partially derived from saline interdune corridors in a manner similar to that which occurs in lunette formation when sediment is blown from saline dry lake floors. However, the abnormally high carbonate content of the Woorinen Formation is less readily explained. Instead, it is likely that a calcareous, suspended load increment, derived during the formation of the extensive coastal dune systems of the Bridgewater Formation, could readily have been carried farther inland and incorporated into the Woorinen Formation, as suggested by Crocker (1946).

Similar pelletal clays to those in the Woorinen Formation were described by Wasson (1983) forming a significant proportion of the dunefields of the Strzelecki and Simpson deserts, and ascribed by him to having blown from saline interdune corridors. Whatever the coastal component, it is clear that the identification of clay pellets in the Woorinen Formation has great significance from the viewpoint of the present-day salinity problem, in that it clearly establishes the distribution and extent of surface salinization in the Mallee during at least six phases of high regional water tables in the past.

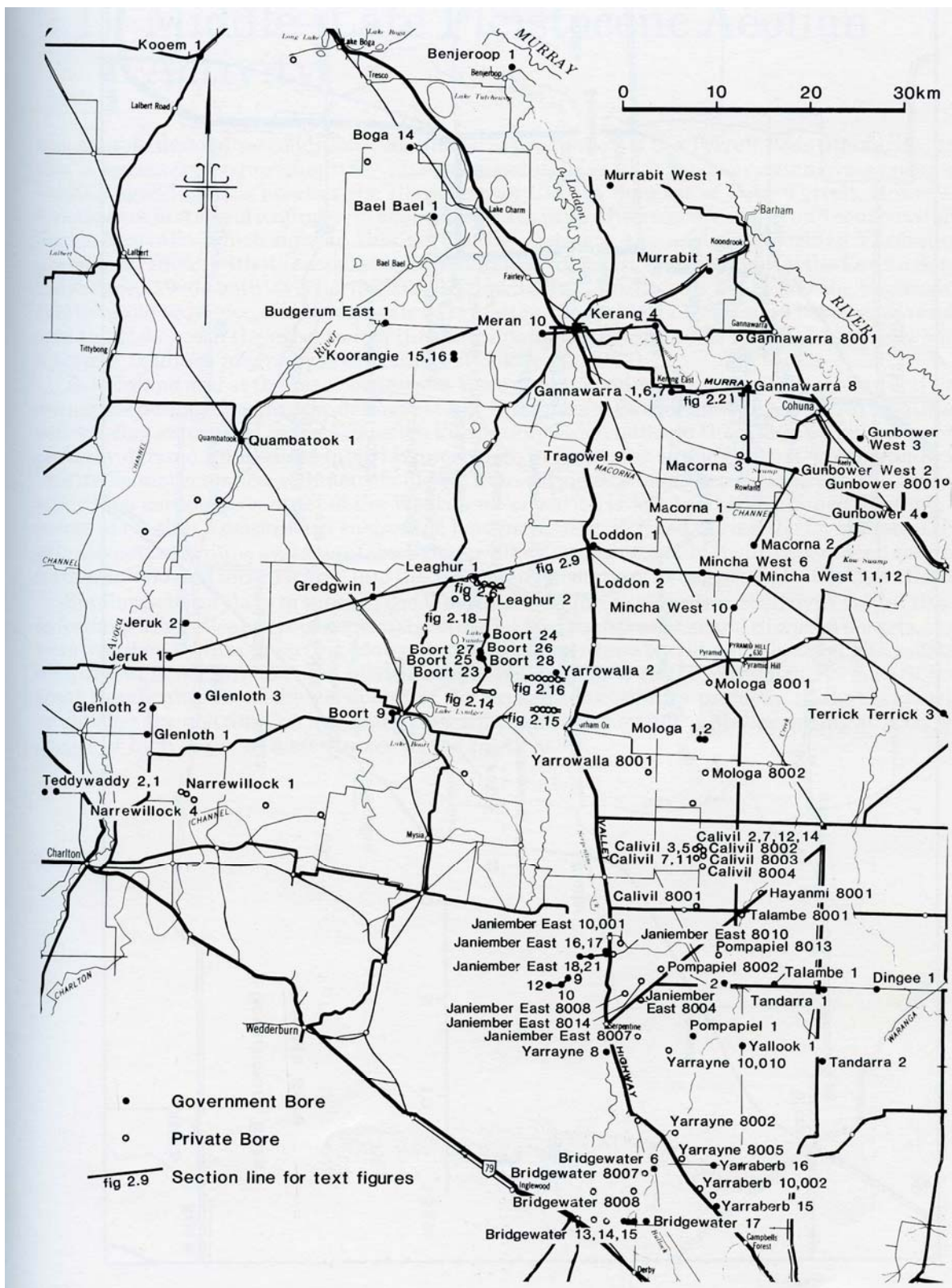


Figure 33 – Location plan of bores and section lines, Loddon and Avoca Plains

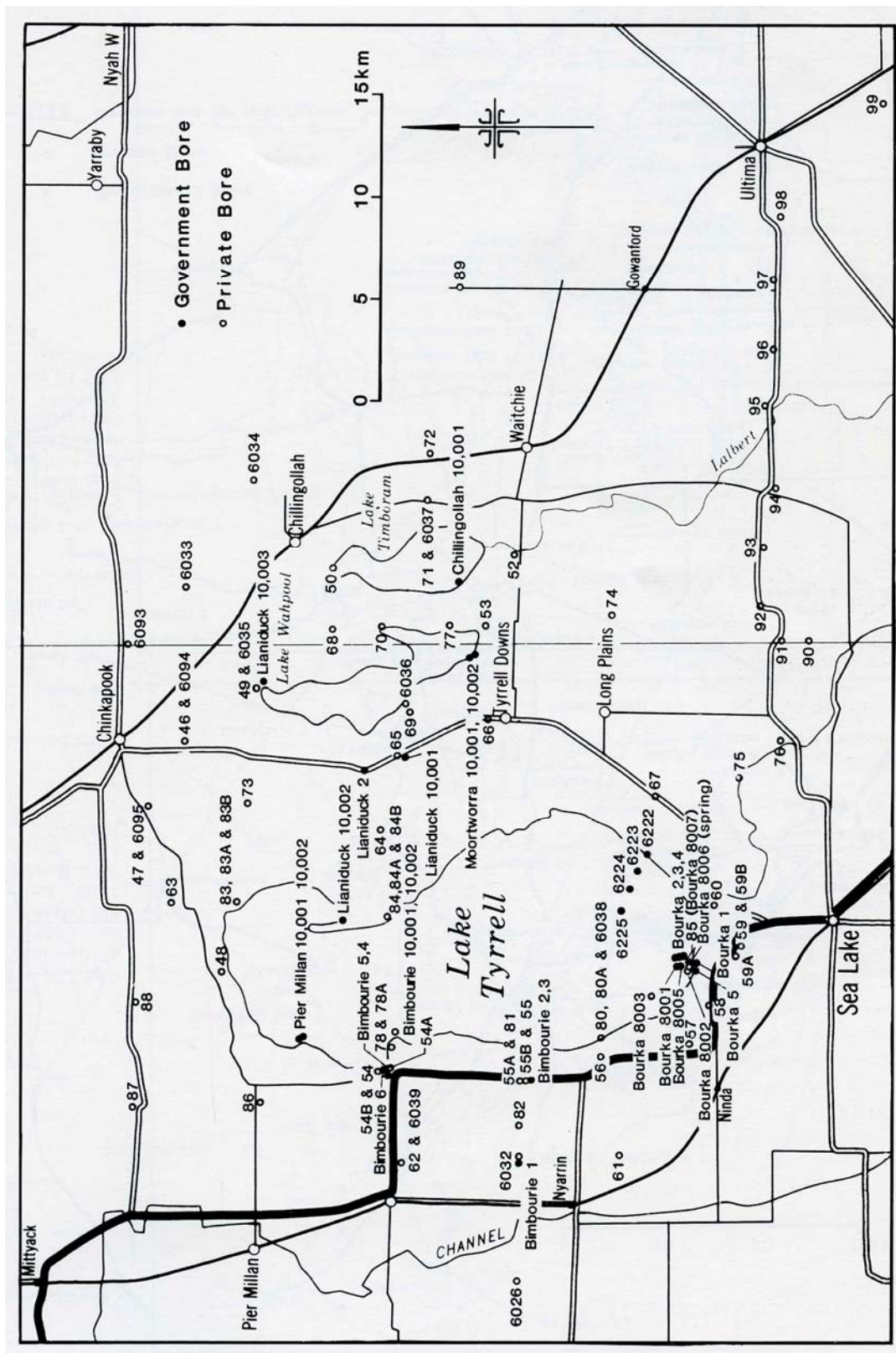


Figure 34 – Lake Tyrrell area – location of groundwater observation bores (after Rural Water Commission)

3. CHRONOLOGICAL FRAMEWORK FOR HYDROLOGICAL CHANGE IN NORTHERN VICTORIA

3.1 Introduction

In any study involving regional groundwater processes, an important perspective can be obtained from an understanding of the palaeohydrology, since past climates have strongly influenced the nature of groundwater/surface water interactions, and hence the aquifer hydrochemistry and the geomorphology. This is the case in the settings of southeastern Australia, where the low landscape relief and the correspondingly low hydraulic gradients result in delicately balanced hydrological regimes, which respond quickly to changes in water budgets. A temporal perspective is also useful when attempting to understand the hydrological changes now taking place, not only across northern Victoria, as discussed in Chapter 4, but across the whole of southern Australia.

The aim of this chapter is not, however, to provide a comprehensive account of the evidence for climatic change in southeastern Australia, but, instead, to provide palaeohydrological data for the broader hydrological study of selected sites and regions which are discussed in this study. Indeed, as is shown here, while lakes may provide an hydrologic record of the conditions existing at a point in the drainage basin through time, there are difficulties in the interpretation of that record to provide an account of the broader hydrologic fluctuations occurring within the overall drainage basin, let alone the climate of the region.

3.2 Previous Work

The lake basins of southeastern Australia reveal a history of major hydrological change spread over the last 50,000 years (Bowler, 1971, 1976; Singh, 1981), which has enabled a broad reconstruction of past climates and environments for this region. To this end, a considerable amount of detailed work was carried out on the lakes and swamps of southwestern Victoria and southeastern Australia (Dodson, 1974, 1975, 1977, and 1979; Bowler 1971, and 1981; De Decker, 1983; and others) and at Lancefield Swamp in south central Victoria (Gillespie et al., 1978). In northern Victoria a number of sites have been dated, although perhaps with the exception of Kow Swamp (Macumber, 1977) and Lake Tyrrell, they have not been investigated (nor, in many instances, do they lend themselves to be investigated) in detail seen at the southern sites. Dated sequences are largely restricted to the western Mallee: along the course of the Wimmera River-Lake Albury (Bowler, 1971; Bowler and Magee, 1978); Lake Wirrengren (Macumber, 1980a Ross, 1981); Lake Wonga (Ross, 1981); from aboriginal campsites further north at Raak Plain (Ross, 1981) - and on the Murray River (Coutts, 1980).

3.3 Lake Tyrrell

3.3.1 Tyrrell Beach

The earliest dated of the Mallee lake sites is at Lake Tyrrell (for locality, see Figures 3.1a and 3.1b), where a high-level beach occurs at an elevation of about 13.5 m above the lake floor. In comparison, the present-day lake reaches a maximum depth of only 0.7 m during even the wettest years. The high beach at Lake Tyrrell is best observed in an erosion gully (Box Gully) at the northern end of the lake (Figure 3.2). The beach sediment is medium- to coarse-grained quartz sand, containing stringers of rolled carbonate nodules and the occasional piece of carbonate encrusted bone; small *coxiella* shells are scattered throughout. The *coxiella* shells suggest that, at the time, Lake Tyrrell had a salinity of between 30,000 and 50,000 mg/l (based on the work of Williams, 1981, who examined the salinity range of *Coxiella striata* in western Victorian lakes). It was significantly less saline than the present lake water, which rarely has a salinity of less than 250,000 mg/l and is commonly over 300,000 mg/l. The former extent of the enlarged Lake Tyrrell is shown in Figure 3.3. At the same time, lakes Wahpool and Timboram would have been incorporated into a single megalake - Lake Chillingollah - extending well beyond the limits of the two present-day lakes. Remnants of Lake Chillingollah are seen today in a string of small groundwater discharge depressions, passing northward from Waitchie to beyond the Towan Plains. On passing landward from the ancient shoreline, the beach sands develop a humic soil cover and grade laterally firstly into a heavily humic-stained clayey sand, rich in small pieces of animal bone, and then into a humic-stained, sandy clay (Figure 3.4). The sequence represents a lateral facies change from a beach environment to a back beach grading into a back swamp environment. The beach is sharply overlain by a reddish brown lunette, which is in turn overlain by a grey lunette (Figure 3.2); the boundary between the two lunette sequences being marked by a well-developed palaeosol and signs of much burning. Charcoal from the base of the lower lunette gave a date of 31,700 $\pm$ 1,140 BP (SUA-559); a second date, also on charcoal, gave an age of 37,500 $\pm$ 2,900 BP (SUA-1492).

There was an initial problem in the interpretation of these dates: the more recent date is stratigraphically from directly above the beach, while the older date is from vertically higher in the section but further back from the shoreline. However, a re-examination of the stratigraphic section suggests that the area of burning associated with the older date, lies within the humic zone developed on the back beach sediments which rises away from the shoreline. It is therefore better correlated with the phase of higher lake levels, which terminated about 32,000 years bp (based on SUA-559, above). There is a probable age correlation between the Tyrrell beach and lacustrine sediments, which contain freshwater mussels beneath a linear dune on the Ouyen Highway about 35 km northwest of Lake Tyrrell; they are dated by Bowler (1980) as being older than 38,000 BP (ANU-1017). Bowler comments that these mussels almost certainly belong to a stage of high lake levels; he states that a true date of 40,000 to 50,000 BP is probably realistic.

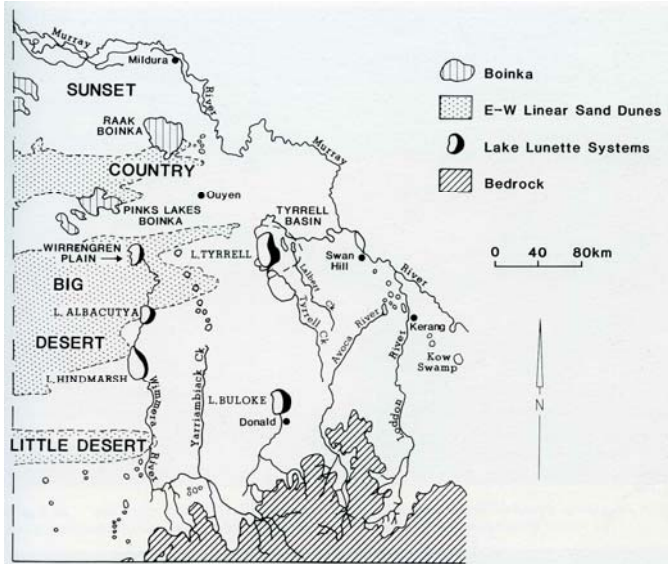


Figure 3.1a - Locality plan for northwestern Victorian sites

3.3.2 Lower Lunette

The lower lunette has a well-developed red-brown earth palaeosol with the following general description:

Depth (m)	Horizon	Description
0.0 - 0.12	'B'	Bright brown (7.5 YR 5/6), fine, sandy clay to loam.
0.12 - 0.60	'C'	Bright reddish brown (5 YR 5/6), light clay passing into a medium to heavy clay, moderate pedal development with prismatic to blocky structure.
0.60 - 1.30		Bright reddish brown to light yellow-orange (5YR 5/6 - 7.5 YR 8/3), light to medium clay, with much carbonate as small nodules, tubules, and as soft, earthy patches.
1.30 - 2.0+		Bright brown (7.5 YR 5/8), light to medium clay, with the occasional carbonate pellet, and carbonate and gypsum tubules and pipes.

Wherever it is exposed around the lake, the surface of the lower lunette commonly shows evidence of extensive burning prior to its being buried under the grey lunette. Whether the burning was natural or whether it was due to human occupation of the Tyrrell shoreline is not known; however, given the high salinity of the lake at the time, it seems that prolonged occupation of the shoreline is very unlikely. There is evidence of at least occasional visits by humans to the lake, as seen by the occurrence of small chert artifacts, pieces of emu shell and burnt clay in one burnt area in Box Gully. A date on charcoal from this site, in the 'A' horizon, was 23,400 +/- 340 BP (SUA-1439). Other dates from the charcoal horizon of the lunette were: 22,000 +/- 730 BP (SUA-588), obtained from the southern end of Box Gully, and 27,780 +/- 730 BP (SUA-783), from the southeastern end of Lake Tyrrell (for locality plan, see Figure 3.1b).

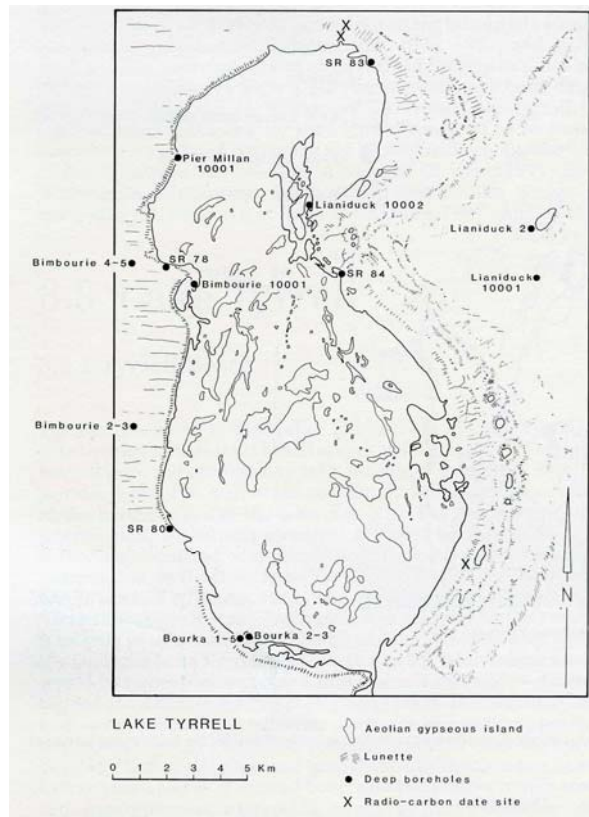


Figure 3.1b – Geomorphology of Lake Tyrrell

3.3.3 Upper Lunette

The lower lunette is in turn overlain by a grey lunette, which also has a well-differentiated soil profile, similar in many respects to that on the lower lunette. A description of the soil sequence is as follows:

Depth (m)	Horizon	Description
0.0 - 0.12	'B'	Dull brown (7.5 YR 5/4), sandy loam to sandy clay loam.
0.12 - 0.40	'C'	Dull brown (7.5 YR 5/3), medium to heavy clay.
0.40 - 0.90		Mottled, light to medium clay, with carbonate concretions which occur as nodules and tubules.
0.90 - 1.60+		Dull brown (7.5 YR 6/3), medium clay, with blocky to prismatic structure.

While no dates have been obtained directly from the grey lunette, similar sediments forming the aeolian islands at Lake Tyrrell have been dated at around 20,000 BP (J.M. Bowler, pers. com).



Figure 3.2a – Box Gully, Lake Tyrrell. The grey lunette (A) overlies a reddish brown lunette (B), which in turn overlies a sandy beach (C). The 'A' horizon developed on the upper lunette is clearly visible (D).



Figure 3.2b – Box Gully (detail) – A, B, and C stratigraphic horizons as above. The pedogenic 'A' horizon developed on the lower lunette is also clearly shown.

3.3.4 Discussion on the Tyrrell Shoreline Sequence

The high lake level recorded in the beach deposits of Lake Tyrrell is in accord with data from elsewhere in southeastern Australia. This indicates that a major wet (lake-full) phase existed from prior to 50,000 years up to just after 30,000 BP. From the semiarid regions, the high lake phase was first recognized by Bowler (1971) at Lake Mungo as existing prior to 26,000 years BP and at Lake Frome before 30,000 years BP (Bowler, 1981; Callen, 1983). Elsewhere in southeastern Australia it was recognized as occurring prior to 25,000 years BP at Lake George (Coventry, 1976), at about 36,000 to 30,000 years BP at Lake Leake (Dodson, 1975), and at Lancefield Swamp (Gillespie et al., 1978) prior to 26,000 years BP.

The change from high lake levels to an environment conducive to lunette formation represents by far the greatest single hydrological change in northwestern Victoria over the last 50,000 years. Indeed, in order to produce a lunette at Lake Tyrrell, after drying of the lake conditions would need to be even more arid (at least seasonally) than they are today. Under the present-day regime, summer water tables fluctuate from being virtually at the surface to about 30 cm below the surface, that is, water tables are too high for lunette formation to proceed. In this respect Lake Tyrrell is similar to many lakes in the Loddon Valley of northern Victoria, for example, Lake Tutchewop (Figure 3.5), where similar dual lunette systems to that occurring at Lake Tyrrell have been previously described (Macumber, in Bowler and Macumber, 1973). The period of soil formation on the lower lunette at Lake Tyrrell dates from about 28,000 up to 22,000, but probably extended until about 20,000 years BP. This period is seen as being one of decreased aridity at Lake Tyrrell compared with the periods both before and after when lunette formation was underway. The presence of a significant vegetation cover at that time is suggested by the extensive burning found wherever the surface is exposed.

3.4 Comparisons with Lancefield Swamp

The severity of the change in hydrology during the drying of the megalakes (Bowler, 1981) is particularly well illustrated at Lancefield Swamp (37°16' S, 144°44' E), a spring-fed swamp in southern Victoria (Gillespie et al., 1978), where a major decline in the spring flow just prior to 26,000 BP led to the partial backfilling of the swamp. Although occurring in southern Victoria, this site is included in this discussion because it clearly demonstrates how the groundwater system can act as a localized buffer to hydrologic change.

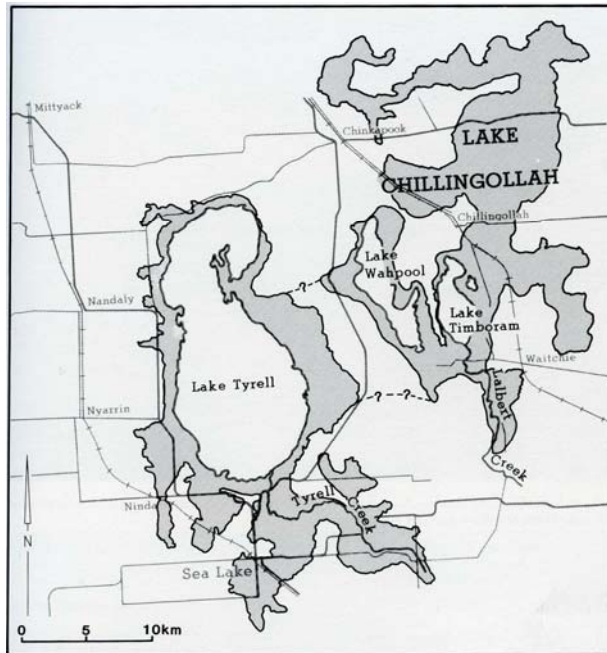


Figure 3.3 - Former extent of the Tyrrell Basin lakes at 40,000 B.P.

The sediments initially deposited in the swamp consisted of gravelly channel deposits, containing bones of extinct megafauna. Dates on charcoal from the channel deposits gave ages of 26,000 $\pm$ 650 BP (SUA-538) and 25,200 $\pm$ 800 BP (SUA-685). The channel-fill sediments pass upward into a densely packed bone bed, with bones of the extinct megafauna, including *Diprotodon* sp., *Macropus titan*, *Protemnodon anak*, *P. cf. brehus*, *Sthenurus occidentalis*, and cf. *Geniornis*. Sections through the deposit suggest that there were at least 10,000 individuals in the swamp. Several quartzite artifacts were associated with the bone bed, including a large quartzite blade. The Lancefield Swamp was groundwater fed and therefore buffered against severe hydrological changes. The site is seen then as having the requirements of a 'last waterhole' under adverse climatic conditions. Whether the animals died as a consequence of eating out the available food resources near the swamp, as suggested by Horton (1978), or whether the swamp became in time increasingly less viable, with perhaps seasonal drying of the swamp as conditions continued to deteriorate, is not known. Whatever the immediate reason for the extinctions, the ultimate cause is seen as being a direct consequence of the major change in hydrology noted throughout southeastern Australia in the late Pleistocene and recorded in the shoreline sequences from Lake Tyrrell. Sections through the Lancefield Swamp are given in Figures 3.6 and 3.7; a complete description of the site is beyond the scope of this study.



Figure 3.4 – Box Gully, Lake Tyrell – grey lunette (A), reddish brown lunette (B), beach sands (C). Humic stained, sandy clay, back swamp sediments are shown in the foreground (D).



Figure 3.5 – Lake Tutchewop lunette, Loddon Valley. The Tutchewop lunette is one of a number of lunettes in the Loddon Valley showing similarities to the Tyrrell lunette system in having a grey lunette (A) overlying a reddish brown lunette (B).

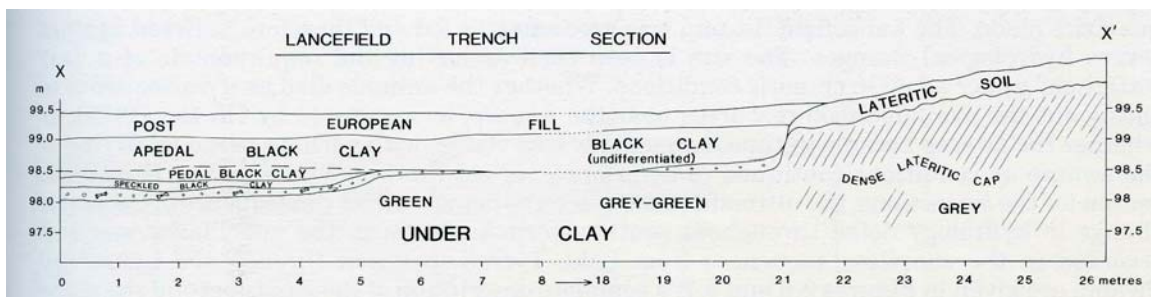


Figure 3.6 – N-S section through part of the spring fed Lancefield Swamp. The section shows 2 stage of incision, with the bone bed lying at the base of the subsequent black clay infill.

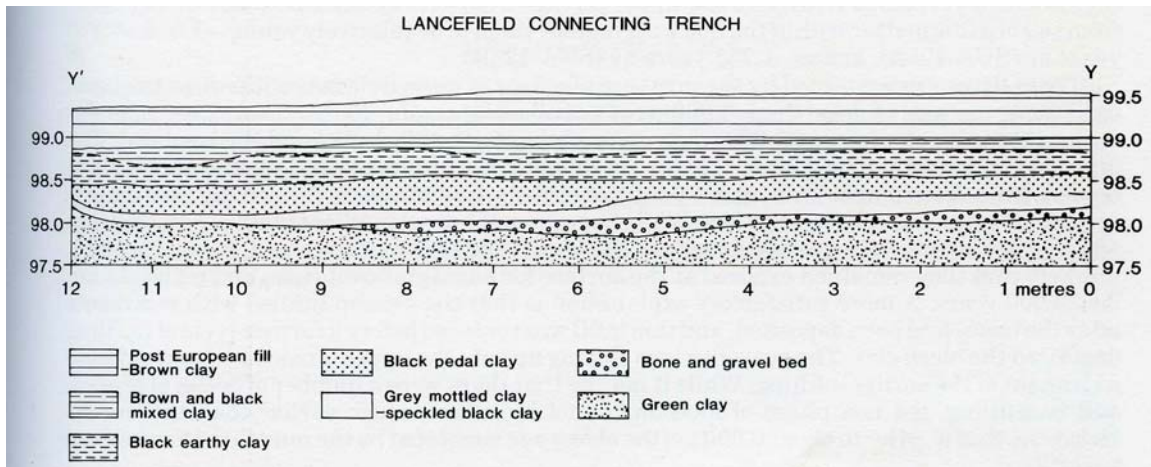


Figure 3.7a – E-W section through part of Lancefield Swamp showing the cut of the bone bed as the green clay base rises on passing down-swamp.

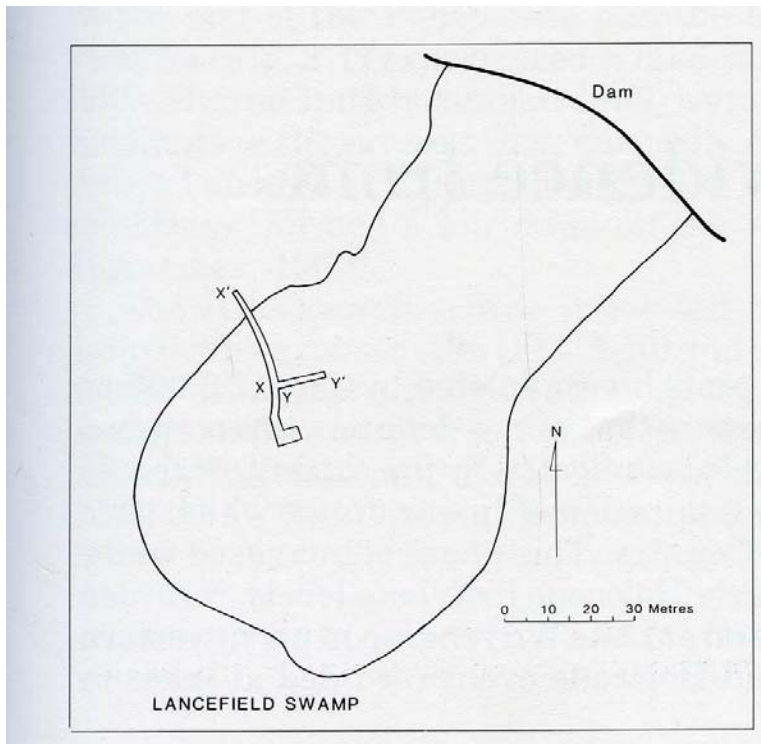


Figure 3.7b – Plan of Lancefield Swamp showing section lines.

The section (Figure 3.6) shows an upper terrace and lower swamp floor, indicating two separate stages in the cutting of the swamp. The bone bed and an overlying speckled black clay are restricted to the lower swamp floor, although gravelly clays of the bone bed form a continuous layer which rises onto the upper terrace where it rapidly cuts out. Carbon dates from soil organic matter within the black clays show them to be relatively young -1,915 $\pm$ 110 years BP (SUA-425A), and ca. 1,735 years BP (SUA-425B).

These dates are supported by the presence of a floor of geometric microliths near the base, indicating an age of less than 6,000 years (Gillespie et al., 1978). Therefore, a major stratigraphic break of some 20,000 years occurs between the black clay and the bone bed. The interval is represented by 0.12 m of black speckled clay, which contains lateritic pebbles and eroded bone fragments. This unit is interpreted as having been formed by low-energy reworking of the underlying bone bed. While bones remain in an excellent state of preservation when in sediment, they quickly deteriorate on exposure to the atmosphere; it is, therefore, unlikely that they remained exposed at the surface for any significant time, and certainly not for 20,000 years. A more satisfactory explanation is that the swamp infilled with sediments after the bones had been deposited, and this infill was removed before a further cycle of infilling deposited the black clay. The gravelly clays sloping up onto the upper terrace are seen as being a remnant of the earlier infilling. While it may be that there were a number of cycles of erosion and backfilling, the last phase of incision cannot have been much earlier than the middle Holocene, that is, prior to about 6,000 BP (the oldest age suggested by the microliths for the final backfill).

3.5 Holocene Period Evidence from Northern Victoria

Although the last major phase of aeolian activity was probably completed by about 16,000 BP (Bowler, 1976), there is evidence in aeolian island sequences that minor deflation has occurred periodically since that time. A period of higher lake levels is indicated by the cliffing of islands at both Lake Tyrrell and the Raak boinka and by the truncation of linear dunes which once transgressed now dry lake beds in the western Sunset Country. This phase of increased water availability is tentatively correlated with a period of early Holocene high lake levels, recorded at Kow Swamp in northern Victoria (Macumber, 1977) and at Lake Wirrengren in northwestern Victoria (Macumber, 1980). It may also relate to a mid-Holocene event recorded at lakes in southwestern Victoria (J. Luley, pers. coin.).

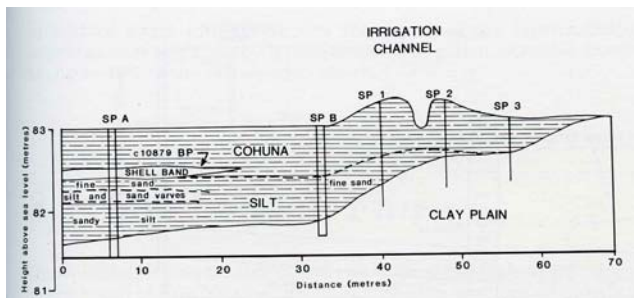


Figure 3.8 - Stratigraphic section through the eastern shoreline of the ancient Kow Swamp (from Macumber, 1977)

Evidence from the lake systems of northern and southern Victoria indicates that in the early Holocene the dry conditions which existed during full glacial times (Bowler, 1976) had passed. In southwestern Victoria lakes began to rise, peaking at about 7,000 to 5,500 years BP (Bowler, 1981). At Kow Swamp, high lake levels existed from about 13,000 to 8,000 BP (Macumber, 1977), and at Lake Wirrengren, from before 7,280 BP (Macumber 1980). The high levels occurring in these Victorian lakes during the early Holocene is in accord with the observations of Street and Glover (1976), who noted similar occurrences in lakes in many parts of the world at this time.

3.6 Kow Swamp

At the end of the Pleistocene and the beginning of the Holocene, Kow Swamp in northern Victoria (Figure 3.1) experienced a lake-full phase, during which time lacustrine silts (Cohuna Silt - Thorne and Macumber, 1969) were deposited at levels well above and in positions outside the limits of the present-day, naturally dry, but artificially filled lake. The lake was fresh and covered an area of 25 km², being 3-4 m deep. Sand, blown from the lake beach under lake-full conditions, formed a low dune on the southeastern shoreline - the Kow Sand (Thorne and Macumber, 1969).

The lake shoreline was inhabited by aborigines, many of whom were interred in the periodically exposed silts (Fig. 3.10) and on the adjacent dune. Preservation of the bones buried in the silts was greatly enhanced by the infusion of the silts with calcium carbonate precipitated from groundwaters under high water table conditions (Macumber, 1977). A radiocarbon date on unionid shells within the Cohuna Silt gave an age of 10,879 $\pm$ 100 BP (ANU-871). This date was from less than halfway down the Cohuna Silt sequence, suggesting that the initial deposition of the silt commenced well before 11,000 BP (Figure 3.8). A pulse of lake expansion at 9,260 $\pm$ 270 BP (ANU-533) led to the incision of previously existing Cohuna Silt and the deposition of a gritty beach containing water-rolled, quartz artifacts (Wright, 1975). The above two dates show that the period for which high lake levels existed at Kow Swamp spanned the Pleistocene-Holocene boundary.

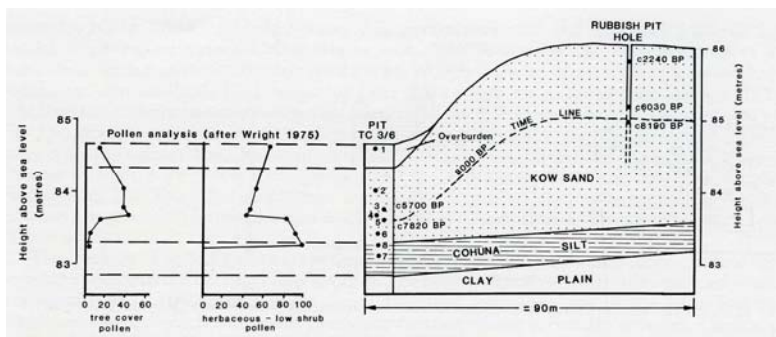


Figure 3.9 - Stratigraphic section through the Kow Swamp sand lunette, with additional chronological and palynological data. Note the rise in tree cover pollen at about 8,000 BP; this results from a fall in water tables following the drying of the lake (from Macumber, 1977).

A series of dates was obtained from the burials within the Cohuna Silt, either directly on bone fragments or on shell or charcoal associated with the burials; these gave a time range which showed that the shoreline was occupied from about 13,000 $\pm$ 280 BP (ANU-1236) to about 8,700 $\pm$ 220 BP (ANU-1038). Despite some differences between several dates from differing materials, the overall range of dates is fairly restricted and conforms to the high lake phase dated from the sedimentary sequences. A further series of radiocarbon dates was obtained from the vicinity of the KS 9 skeleton, which was buried in the nearby Kow Sand dune, and the grave covered by further dune accretion. Dates on KS 9 are 9,300 $\pm$ 220 BP (ANU-619B) from bone apatite; 9,590 $\pm$ 130 BP (ANU-532) from charcoal; and 8,080 $\pm$ 420 BP (ANU-531) also on charcoal (Macumber, 1977).

Additional radiocarbon dates from a vertical sequence in the Kow Sand enabled dune sedimentation rates to be calculated. These showed a sharp fall off in the rate of dune accretion at about 8,000 BP suggesting that the beach was no longer contributing sand to the dune and that the lake had dried up (Macumber, 1977). This event coincided with the appearance of large amounts of tree pollen in the dune sediments (Figure 3.9). The sudden reduction in dune sedimentation rate when coupled with the introduction of the tree pollen at 8,000 BP was interpreted (Macumber, 1977, 1978a) as indicating a fall in regional water tables accompanying the drying of the lake. From 8,000 BP, dune accretion rates remained low and the tree pollen

content high, suggesting that no comparable wet period to that experienced in the early Holocene occurred at Kow Swamp until the coming of Europeans and the artificial filling of the lake.

The Kow Swamp situation is of special interest to present-day environmental studies of hydrologic instability in the Loddon Plain, in that it represents a reversal of the modern trends wherein high saline water tables have come about in response to large-scale irrigation with the consequent destruction of the tree cover. In the case of early Holocene Kow Swamp, a reduction in water availability saw the drying of the lake, a fall in water tables and the appearance of trees (note Figure 3.9, where tree pollen increases from almost nil prior to 8,000 p to 40% of the pollen, soon after).

3.7 The Loddon Fan

One additional date for this period comes from the Loddon fan, situated on the Loddon Plain to the south of Kow Swamp (Figure 2.23). The fan was formed during a phase of higher flow in the Loddon River than at present, its formation being seen as coinciding with high water tables under the Loddon Plain (Chapter 4.13). Charcoal from within the channel sands (Figure 3.11) of a Loddon river distributary gave an age of 8,790 $\pm$ 120 BP (SUA-1320).

Additional palaeofeatures of the fan are the traces of a number of ancient springheads and regulated streams, which have their origin in groundwater outflow from the deeply buried regional aquifers during past phases of high groundwater pressures. The stream systems which arise in this manner are strikingly different in morphology to the distributary systems developed on the normal Loddon River system or its major distributaries. This is discussed in greater detail in Chapter 4.

3.8 Lake Wirrengren

Further evidence for early Holocene high lake levels in northern Victoria comes from the Wirrengren Plain (for locality, see Figure 3.1), the largest of a number of now dry former lake beds at the terminus of the Wimmera river. Clay lunette sequences are developed on some of the smaller lakes, for instance, Lake Agnes, and these probably correlate with the last major phase of aeolian activity at about 20,000 to 16,000 BP (Bowler, 1971, 1976). An earlier, outer lunette, encompassing the entire lake complex, can be seen in section at Kelly's Lookout, where it contains *Coxiella* sp. shells, indicating a brackish to saline environment. However, no age can be placed on this event.

At the northeastern edge of the Wirrengren Plain, an 11 m high, strandline deposit contains a freshwater molluscan fauna, including *Plotiopsis balonnensis*, *Gyraulus meridionalis*, *Corbiculina australis*, *Physastra gibbosa*, and *Velesunio ambiguus*. The shells occur in distinct bands (Figure 3.12), with *Plotiopsis* making up about 95% of the population. *Plotiopsis* is a warm water species, which occurs in the lower parts of the Murray-Darling River system (Smith, 1978); it is not known from the Wimmera River today (Smith, pers. com.). Carbon dates on the shells gave an age of 7,460 $\pm$ 120 BP (SUA-763), (Macumber, 1980). High-level deposits containing shells were also found at Millers Tank on the western edge of the Wirrengren Plain, at levels of between 7.1 m and 8.3 m above the general level of the former lake floor.

Shell deposits were also noted at a low, 1 m high, inner strandline on Lake Wirrengren and dated at 6,980 $\pm$ 110 BP (SUA-764); however, in this instance, *Plotiopsis* was absent and instead the shells were predominantly *Corbiculina*, with an occasional *Velesunio* and *Gyraulus*. The position and height of the deposit suggest that it represents a drying of the lake at about 7,000 BP. Therefore, while there are no data on when high lake levels at Lake Wirrengren commenced, there are fairly conclusive data which suggest that the lake dried up about 1,000 years after the drying of Kow Swamp.



Figure 3.10 - Kow Swamp 17 grave in the Cohuna Silt. The burial was dated at 11,350 $\pm$ 160 years BP (ANU - 1235) on unionid shells placed in the grave. Preservation of the skeleton was enhanced by the infusion of the Cohuna Silt by CaCO_3 of groundwater origin.

3.9 Raak Plains

One further relevant date, indicating wetter conditions than now, comes from the Raak Plains, where *Velesunio* sp. shells from an aboriginal campsite (Figure 3.13) were dated at 7,650 $\pm$ 110 BP (SUA-766), (Ross, 1981). This site lies close to the western edge of the Raak boinka and consists of a large number of artifacts and hearths exposed on the floors of two extensive blowouts in linear dunes isolated from the main linear dunefield further west by lateral expansion of the boinka. The site is just one of many dozens scattered across the Raak boinka; however, in almost every other case occupational evidence consists of relatively few artifacts and/or pieces of ironstone, suggesting that these were not ongoing occupational sites like the dated site. At only one other site (on the eastern edge of the boinka) was there any significant accumulation of *Velesunio* shells.

The presence of *Velesunio* indicates a nearby freshwater source, causing Ross (1981) to suggest that some of the salinas must have held fresh or only slightly brackish water at this time. However, the entire area is underlain by groundwater brines which are in direct contact with the salina floors (Macumber, 1980). It is difficult, therefore, to see how any lake in this environment (Figure 3.13) could be sufficiently fresh to support human occupation, even with the higher water budgets existing in the Mallee at the time.

An alternative explanation offered here is that the site was similar to other occupation sites in the northern Mallee away from the river system, that is, it was situated on a dune soak that existed under a favourable climatic regime. Similar soaks, containing evidence of ongoing occupation in the late Holocene, are at Scorpion Wells and Ross Springs, at the northern end the big Desert. At these sites, fresh water is perched above a clay layer at the base of the dune 41 the water is replenished by infiltration of local rainfall through the surrounding dune sands. Chemical analyses from a number of such freshwater aboriginal soaks, forming an eastwest line across the Big Desert, are given in Table 3.1.



Figure 3.11 - Channel sands deposited by an ancestral Loddon River distributary at the head of the Loddon fan, near Serpentine. A radio-carbon date from charcoal within the sands gave an age of 8,790 $\pm$ 120 years BP (SUA-1320).

Table 3.1 Aboriginal Springs and Soaks - Big Desert and Sunset Country

Big Desert Springs	TDS	Cl	HCO ³ +CO ₃	SO ₄	NO ₃	Ca	Mg	Na	K	SiO ₂
Sandbucket	111	20	41		3	3	1	29	3	10
White Springs	345	72	98		14	11	7	79	8	54
Ross Springs	1347	411	183		17	36	50	337	10	2
Scorpion Wells	115	47	11		8	13	3	17	2	14
Little Billy Soak	81	4	27		7	4	2	14	3	19
The Springs	79	8	9		15	5	2	13	2	23
Sunset Country Springs										
Carters Soak	1045	223	419	10		52	11	233	8	89
Boltons Soak	215	21	10			12	3	46	3	18
Rock Holes Soak	209	25	31		68	19	3	13	25	25



Figure 3.12 – Shell bands (darker colours) eroding out of a sand sequence on a high level strand line of the ancient Lake Wirrengren. The shells gave a radio-carbon age of 7,460 ± 120 years BP (SUA – 763).



Figure 3.13 – Raak 1 and Raak 2 archaeological sites in the Raak boinka. Unionid shells from occupation areas in the Raak 2 site date at 7,650 ± 110 years BP (SUA – 766). Note the present day saline landscape, which makes it unsuitable for occupation.

At Raak the dunes directly overlie Blanchetown Clay which often occurs within a metre of the lake floor. A situation may have arisen at the site whereby a freshwater lens existed, overlying saline groundwaters, as is commonly found in coastal dune systems. While this situation could be maintained within a dune 'perched' aquifer system, it is unlikely that it would be maintained for any length of time in a shallow lake directly overlying groundwater brines. The case for a freshwater lake rests entirely on the presence of *Velesunio* shells, and must therefore be asked why these shells may not have been imported from an expanded freshwater Hattah Lake system, lying immediately to the east of the Raak boinka.

3.10 Lake Correlation - A Discussion

The late Pleistocene lake-full phase (the megalakes of Bowler, 1981) observed at many places in southeastern Australia is represented at Lake Tyrrell by the 13.5 m high beach. This phase lasted at Lake Tyrrell until about 32,000 BP, when the switch from very high lake levels to seasonally dry lakes led to the development of the lower lunette. This event represents by far the biggest change in surface water budgets at any time over the last 50,000 years. It is reflected in the demise of large numbers of megafauna at Lancefield Swamp in southern Victoria, at about 26,000 BP.

A later lake-full episode, documented in northern Victoria, commenced at Kow Swamp at about 13,000 BP. There are as yet no dates showing when this phase commenced in the Mallee, further west; however, the wider evidence for the amelioration of the arid high glacial conditions by the start of the Holocene suggests that it may have commenced at about this time. This lake-full phase is of much smaller magnitude than that recorded earlier in the late Pleistocene; however, the data from Kow Swamp and from the Sunset Country suggest that higher lake levels than at present and the accompanying higher regional water tables were being experienced across northwestern Victoria in the early Holocene Period. The lake-full stage lasted until about 8,000 BP at Kow Swamp and 7,000 BP at Lake Wirringren. At neither site has evidence been found of a comparable wet phase at any later stage during the remainder of the Holocene.

In summary, therefore, the lakes of northern Victoria which were examined for this study show a record of climatic fluctuation, which broadly agrees with that occurring throughout southeastern Australia during late Pleistocene and early Holocene times. It is only when sites are compared in detail that real differences begin to appear, for instance, the peaking of Lakes Keilambete, Bullenmerri and Gnotuck at 7,000 to 5,500 BP (Bowler, 1981) at a time when Kow Swamp had dried out and Lake Wirringren was drying. However, these differences are not surprising, given the very significant differences in lake regime that may exist even within one drainage basin. Indeed, a lake may be considered as the surface expression of the hydrological equilibrium that exists within a drainage basin at a point in that basin. Here, the downbasin flow is taken as having two components - a groundwater component and a surface water component, between which there are varying degrees of interaction.

The way in which a lake responds to changes in water budget within the basin depends upon its position in the basin with respect to the varying hydrological thresholds and the physical parameters of the basin. In some instances, small increases in water budget may have large effects on the lake and, hence, on its resultant sedimentary record; in other instances, large changes in water budget may have little or no effect on an individual lake level. It also follows that lakes in one part of a drainage basin may best record wetting up events, while lakes in another part of the basin may best record drying events. A modern day example occurs on the Wimmera River in northwestern Victoria, which has the permanently wet Lake Hindmarsh in the south and the dry lake bed of the Wirringren Plain in the north (downstream). When the Wirringren Plain is underwater, as at 7,400 BP, all lakes along the Wimmera River are full and regionally wet conditions prevail. However, any record of dryness at Wirringren Plain can only be applied to that lake alone, since it tells nothing about the upstream parts of the system. Similarly, a record of wetness from Lake Hindmarsh tells nothing of regional wetness, whereas a dry Lake Hindmarsh implies regional dryness. It is clear therefore, that to disentangle the various signals recorded in a lacustrine sequence, it is necessary that the sequence should be interpreted within the perspective of the overall hydrology of the drainage basin. For instance, one very important threshold that exists on drainage basins in northern Victoria is that which marks the hinge line between regional groundwater recharge and regional groundwater discharge. Freshwater influent lakes are normally found upbasin of the hinge line, while brackish and saline discharge lakes commonly occur on the downbasin side of the hinge.

Examples of widely differing lacustrine regimes in a single drainage basin can be seen on

passing downbasin along the Wimmera River: the lake regimes range from permanent freshwater (Lake Hindmarsh); semipermanent fresh to brackish on drying (Lake Albacutya); dry lake being filled only rarely (Lake Wonga and other lakes of the Wyperfeld area); dry lake but supporting an extensive forest of the native pine, *Callitris preissii* (Wirrengren Plain); permanent salt lakes drying seasonally to give a salt crust (lakes of the Pink Lakes boinka).

The gradual drying of the Wimmera River as it passes inland largely determines the character of the lake regimes up to the Wirrengren Plain; however, at this point the influence of the underlying groundwater system is felt for the first time. At Wyperfeld, the regional water table is about 10 m below ground surface but rises to only about 4 m below the surface at Wirrengren Plain, the terminal (dry) lake basin on the Wimmera River. The groundwater is nonsaline at this point, and it has been suggested (Macumber, 1980) that the *Callitris* forest at Wirrengren Plain owes its existence to the shallow groundwater. The rising water tables reach the surface further north, near Underbool, and beyond this point groundwater-fed lakes are saline, reflecting in part the strong evaporitic regime, which concentrates the salts being discharged from the aquifer systems. Therefore, despite the general overall evenness of the climatic regime, in northwestern Victoria there is a wide range of lake regimes. It is speculated that should the Wimmera system be studied in the future, say, 20,000 years hence, with a view to determining the climate in southeastern Australia in the year 1983, then there could be proposed almost as many different climates as lake systems studied.

Exact correlations become even more uncertain where lakes are exposed to the influence of rising sea levels, as in the case of coastal sites. Here, the effects are not simply the direct effects of increased rainfall, etc.; major changes also occur in the whole basin hydrology as groundwater systems readjust to newly established base levels and develop their own new thresholds.

It follows that the significance of the palaeohydrology of an individual lake must be interpreted in terms of the overall regional hydrology of the entire drainage basin.

In conclusion, therefore, the individuality shown in the behaviour of lake systems to hydrological change is an important characteristic of lakes, and for this reason no generalized correlation table is included in this study. Instead, when necessary, each part of the basin is considered on its own merits, and where required, reference is made to the palaeohydrologic data from that part of the basin, rather than to any Australia-wide monolithic generalization.

4. RIVERINE PLAIN AND ADJACENT HIGHLAND CATCHMENTS

In the Riverine Plain of northern Victoria, regional groundwater flow passes downbasin, from recharge areas within the highlands and on the plain, towards regional discharge areas, mainly in the Mallee (Chapter 6), but also occurring in the Loddon Valley in the westernmost part of the Riverine Plain (Figure 4.1). The main regional aquifers are the Calivil Formation, and the Warina Sand - the coarse unit in the Renmark Group. The aquifers are essentially valley-fill deposits, which infill valleys that are incised into pre-existing units: the Renmark Group cuts into bedrock, while the Calivil Formation valleys cut across either bedrock or Renmark Group sediments (Chapter 2).

This section examines the interactions between groundwater and surface water systems in the Riverine Plain.

4.1 Groundwater Hydrology of the Loddon Basin

4.1.1 Catchments

The main catchment areas for the Loddon drainage basin (Figure 4.2) lie on the northern slopes of the east-west trending Western Highlands, which make up a western physiographic province of the Great Dividing Range in Victoria. The highlands are comprised predominantly of Palaeozoic rocks and Tertiary basalts, which together form little more than a broad flat low divide - the Ballarat plateau (Thomas, 1952). The outpouring of basalts which form the Ballarat plateau infilled existing, deeply incised Tertiary valleys and modified the drainage system so that it led to a major northward shift in the Tertiary divide. Much of the intervening area, which was formerly part of the Loddon drainage basin, now drains southward to the Southern Ocean. The surface drainage basin of the Loddon River is, today, smaller than its late Tertiary equivalent (Thomas, 1952). It is also smaller than the present-day groundwater basin, which is substantially the same as it was in Tertiary times.

4.1.2 Groundwater in the Loddon Valley Lead

Prior to the drought periods of 1967-68, little was known about the deeper flow systems which pass across the Riverine Plain in Victoria. Within the highlands, however, detailed drilling across the major valleys during the latter part of the 19th century and early 1900s had supplied a very large volume of information on the morphology of the valleys. Mining operations had shown that vast quantities of water existed within the auriferous lead systems.

Possibly the best documented material on the characteristics of the groundwater in the highland valleys comes from the records of the relatively late phase of deep lead mining activity, which took place in the first decade of the 1900s in the central Loddon Valley near Maryborough. During this period, there was established along the Loddon Valley lead system and its tributaries a number of large-scale alluvial mines, operating on the Berry-Moolort trunk lead and on two important tributaries - the Majorca-Carisbrook tributary and the Bet Bet Valley system, the latter being the eastward continuation of the trunk Avoca Valley system.

In the Annual Report of the Department of Mines for 1905, the pumping capacity of the Loddon Valley Gold-fields Company Ltd, from a single shaft, is given as 320 Us. The Charlotte Plains Consolidated Gold Mines Co. Ltd pumped at 320 - 470 l/s, and its plant, established as the Deep Leads Electric Transmission Co., used 645 tonnes of firewood each week, "so that in a year a forest is consumed by this plant alone" (Australian Mining Standard, 1906). When pumping commenced, in 1902, the water level was 40.2 m below the surface in the New Havillah and Charlotte Plains lease, and after 12 months pumping the level had fallen to only 52.7 m (Australian Mining Standard, 1903). During this period, Chalks No. 3 Consolidated Co. pumped at about 260 l/s. In the immediate vicinity was the Moolort Gold-fields Co. Ltd, which pumped 22,680 m³ during the period January to August 1905; downstream from Carisbrook, the Victoria Deep Leads Ltd pumped at an average of 200 l/s for 1906. In 1905 the Grand Duke and North Duke mines pumped at 220 l/s.

These large volumes of water demonstrated that the lead system was a major drain for groundwater passing northward from the highlands.

Detailed investigation of the downstream continuation of the Loddon Valley lead commenced in 1968, and the subsequent drilling program traced the well-defined valley-fill system northward towards the Murray River, where it was found to junction with a Murray Valley system prior to entering the sea.

The lead sequence, consisting predominantly of sand and gravel, was traced northward beneath the Loddon Plain, where it was designated the Calivil Formation (Macumber, 1973). It is named after the parish of Calivil where the aquifer system in the gravel is now extensively used for

irrigation purposes. The type section is in the Mines Department Calivil No. 2 bore, where gravel and sand occur from 77 m to 95 m; it is incised into and overlies the carbonaceous Renmark Group. It is, in turn, overlain by the Shepparton Formation, which forms a varying permeable aquitard. The gravel and sand are clean throughout the entire length of the system, as suggested by their exceptionally high permeabilities, measured from Bridgewater in the south to Gunbower in the north (Table 4.1).

Measurements of the permeabilities of the Calivil Formation were carried out at Bridge-water, Calivil, Mincha West, South Cohuna and Yando by means of pump tests, using a discharging bore and one or more observation bores. Constant discharge tests were carried out, and the results were analyzed by using the Theis type-curve solution for radial flow in an infinite leaky aquifer (Theis, 1935; Jacob, 1940).

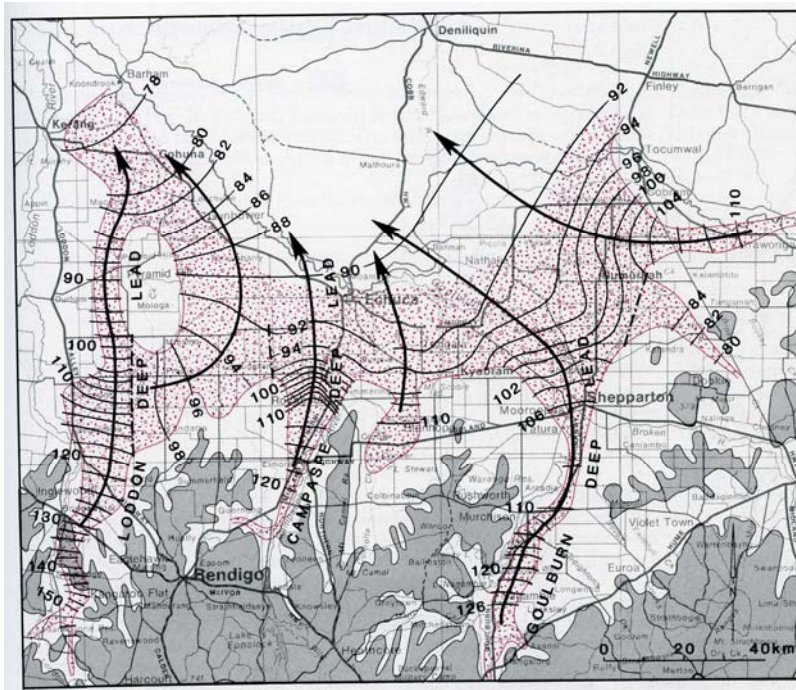


Figure 4.1 - Regional groundwater flow direction in the Riverine Plain in Victoria (from Tickell and Humphrys, 1986)

TABLE 4.1 - Permeability of the Calivil Formation-Loddon Valley

Bore	Transmissivity	Hydraulic Conductivity
Bridgewater 15	$5.2 \times 10^3 \text{ m}^2/\text{day}$	85 m/day
Calivil 2	$8.9 \times 10^2 \text{ m}^2/\text{day}$	49 m/day
Mincha West 12	$6.6 \times 10^2 \text{ m}^2/\text{day}$	43 m/day
Gunbower West 2	$4.3 \times 10^3 \text{ m}^2/\text{day}$	179 m/day
Boort 23	$1.6 \times 10^3 \text{ m}^2/\text{day}$	94 m/day

On the Loddon Valley deep lead the hydraulic conductivity at Bridgewater is 85 m/d but falls to 49 m/d at Calivil. Gunbower West is on a Murray Valley system, while Boort 23 is on the Powlett Plains lead.

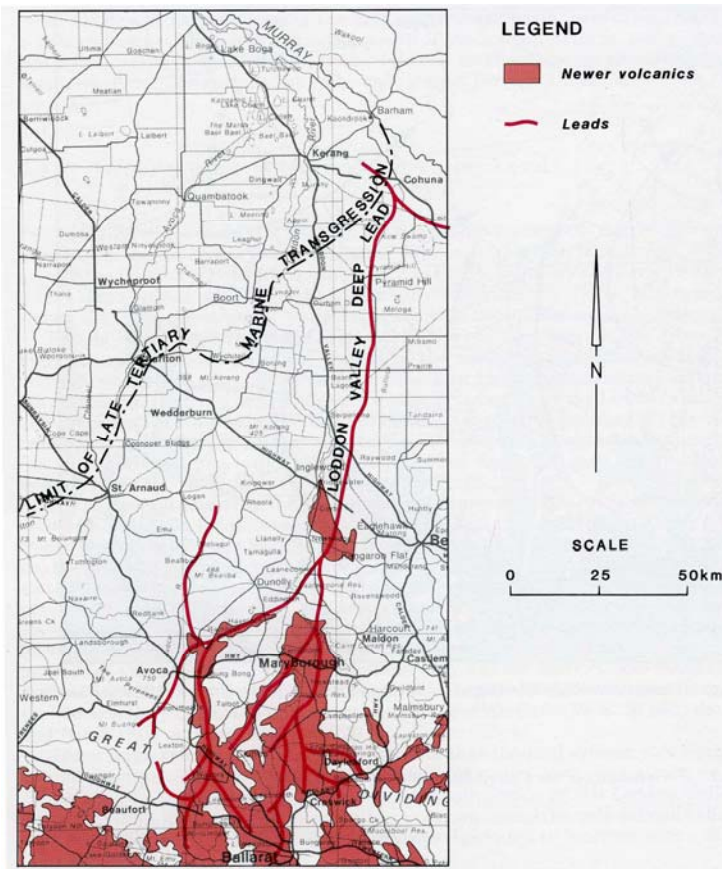


Figure. 4.2 - Position of late Tertiary Loddon Valley drainage system

Ongoing investigations (Macumber, 1978a, 1978b, 1978c, 1983; Tickell and Humphrys, 1980) have shown that the Calivil Formation is the most significant regional aquifer in Victoria for transporting groundwater from highland catchments, across the Riverine Plain, towards regional groundwater discharge zones further downbasin, both on the plains and in the Mallee.

4.1.3 Groundwater in the Parilla Sand

Of the three members of the Parilla Sand developed on the Loddon Plain, the Kerang Sand is the major aquifer. In general, water in the Parilla Sand is fairly saline (Table 4.2), as is the case elsewhere in the Murray Basin. The following examples are given for bores situated in marine sediments east of the Avoca River, on the Gredgwin Ridge and Loddon Plain.

TABLE 4.2 - Saline waters from the Parilla Sand

Bore	Locality	Depth (m)	TDS (mg/l)	Unit
Boort 21	Boort	47.5 - 48.2	23,000	Kerang Sand
Boort 24	Yando	42.7 - 51.8	16,150	Kerang Sand
Boort 8001	Canary Is.	87.8 - 88.4	14,000	Kerand Sand
Budgerum East 1			22,033	Parilla Sand
Budgerum East 8001	Quambatook	18.3 - 39.6	20,000	Parilla Sand
Gredgwin 1	Barraport	41.5 - 43.9	30,040	Parilla Sand
Loddon 2	Macorna	52.1 - 70.1	16,585	Kerang Sand



Figure 4.3 – Lake Kelly on the lower Loddon Plain – a groundwater discharge lake where halite was once regularly harvested as seen in this photo taken in 1984. The lake is now used as an evaporating basin for the disposal of saline water from the irrigation district.

The above high salinities are typical of those occurring in the Parilla Sand; however, an important exception to the above situation occurs in the Kerang-Cohuna district, at the mouth of the Loddon Valley lead where it meets the marine sequences after junctioning with a fresher Murray Valley system. Here, the Kerang Sand directly overlies the Calivil Formation. There is no restriction on groundwater which passes out of the Calivil Formation into the Kerang Sand, and this causes flushing of any pre-existing saline waters and their replacement by better quality waters. The flushed zone extends from west of Cohuna to Kerang. However, the path taken by through moving groundwater beyond Kerang is not known due to lack of deep borehole information. Table 4.3 lists a number of bores in which the transmission of better quality water from the Calivil Formation to the Kerang Sand can be seen.

Regional groundwater discharge occurs in the lower regions of the Loddon Plain where groundwater-derived evaporite deposits of gypsum and halite have been harvested (Figure 4.3). Virtually the whole of the central and lower Loddon Plain lies within a major irrigation district, where salinity problems appeared soon after irrigation commenced late last century; today, much of the plain suffers from high water tables and salinity.

TABLE 4.3 - Salinity in the lower Loddon Plain

Bore	Locality	Depth (m)	TDS (mg/l)	Comments
Gunbower 8001	Gunbower Is	93 - 101	3,550	Calivil Fmn
Gunbower West 2	Orfield	79 - 103	4,512	Calivil Fmn
Gunbower West 3	Orfield	88 - 106	4,326	Calivil Fmn
Macorna 3	Hird's Swamp	58 - 62	4,400	Kerang Sand (overlies Calivil Formation)
Gannawarra 1	Kerang	73 - 76	5,512	Kerang Sand
	Agricultural Research Farm	94 - 99	6,295	Calivil Fmn
Meran 10	Wandella Lake West	49 - 73	5,886	Kerang Sand
	Kerang			

4.1.4 The Influence of Groundwater Flow on Surface Processes

Monitoring the pressure levels within the Calivil Formation aquifer has shown it to be highly sensitive to abnormally wet years (as in 1973). During these periods, pressure levels may rise substantially, and this may in turn influence various processes at the surface. The following section discusses the delicate hydraulic balance that exists within the Calivil Formation aquifer and its influence on the geomorphology and salinity of the Riverine Plain in Victoria.

4.1.5 Regional Groundwater Flow

The effect of water table configuration on regional groundwater flow was demonstrated initially by Toth (1962, 1966). In considering the effect in an area of constant gentle regional slope, he concluded that, given a homogeneous medium, groundwater flow is essentially horizontal, with recharge concentrated at the upstream end of the recharge area and discharge at the downstream end of a discharge area. The hinge line separating the areas is midpoint, except where a major valley is present; in this instance, recharge covers most of the upland area, and discharge is concentrated in the valley. In recharge areas, therefore, a component of the groundwater flow is directed downwards, away from the water table, while, in the discharge areas, there is an upward flow component towards the water table. Despite certain limitations on the use of this steady state flow concept under the transient conditions brought on by water table fluctuations, when used on a regional scale, as is the case in the Loddon Valley, it provides a useful framework by which to examine changes in what is, in reality, a dynamic equilibrium within the flow system.

By the use of digital computer solutions of mathematical models, Freeze and Witherspoon (1966, 1967) examined steady state regional groundwater flow. With this method, they put forward a series of potential nets to cover regional groundwater flow in a number of hypothetical basins, which have diverse lithology and permeability variations within the sequence. An amplification of the Freeze and Witherspoon papers appears in Freeze's Theoretical Analysis of Regional Groundwater Flow (1969a). On factors controlling the distribution of recharge and discharge areas, Freeze concluded that in a simple two-layer system in an area of constant regional slope, the hinge line occurs at the midpoint of the basin. The effect of relative permeabilities of aquifer to aquiclude is that with an increase in permeability ratio the hinge line moves upslope; also large permeability ratios are conducive to large discharge areas, in a simple system where a subsurface highway exists (Freeze, 1969a). Of relevance to the situation on the Loddon Plain where the Calivil Formation lenses downbasin from a thickness of about 60 m at Bridgewater to about 15 m at Calivil (see below), is the comment by Freeze (1969a) on the particular importance of the position and shape of a basal high conductivity layer to the establishment of discharge zones within the drainage basin. He observes that a pinchout of the aquifer will produce a discharge area part way down the basin, with surface topography playing no part here in the positioning of the discharge zone.

Toth (1966) and Meyboom (1966a) demonstrated the practicability of mapping recharge and discharge in the Canadian prairie by means of field observations, while Freeze (1969a) listed field techniques for delineating regional groundwater flow regimes. These included:

1. Piezometric observations
2. Hydrochemistry
3. Surficial evidence
4. Low-flow method

Each of these techniques has been used to varying degrees in the study of regional groundwater flow in the Loddon Basin. Of particular relevance to this study is the generalized plot of depth versus depth to the static water level (Freeze, 1969b; Freeze and Cherry, 1979), which shows the fields expected for groundwaters from recharge and discharge areas (Figure 4.4). One further observation (Freeze and Cherry, 1979) is that, while salinity generally increases along the flow

path, water from the recharge areas is generally fresh, and water from the discharge areas is often relatively saline.

4.1.6 Geological Controls on the Potentiometric Surface

In 1968, the establishment of a series of observation bores was begun in order to monitor fluctuations in the deep groundwater system beneath the Loddon Plain. Permanent Leopold-Stevens water level recorders were established on bores as they were drilled, and those not so equipped were monitored monthly by hand. Eight observation bores were established tapping the Calivil Formation on the main lead valley between Bridgewater and South Cohuna (Figure 4.6a).

The potentiometric surface that was obtained prior to 1973 was based on bores along the infilled lead valley, both within the highlands and on the plain (Figure 4.6b). It was a gentle curve, some 24 m below the surface at Laanecoorie, but rose to only 12.5 m below the surface at Bridgewater, and reached the surface at Calivil, in the central Loddon Plain. Beyond Calivil, bores tapping the Calivil Formation either flowed or were close to flowing (Figure 4.5). This upward flow component indicates that the Loddon Plain beyond Calivil is an area of regional groundwater discharge (cf. Freeze and Cherry, 1979).

Therefore, prior to 1973, the hinge line between the regional groundwater recharge and discharge zones was in the vicinity of Calivil, that is, about halfway across the Loddon Plain. An explanation of the development of the discharge zone lies in the downbasin wedging of the Calivil Formation (Figure 4.7b), with a consequent reduction in aquifer thickness from about 60 m at Bridgewater to about 15 m in the Calivil district (Macumber, 1978a). The reduction in cross-sectional area of the Calivil Formation within the Loddon trench is accompanied by a decreased hydraulic gradient, which together result in an increasing groundwater loss into the overlying sediments and, at times, to the surface.

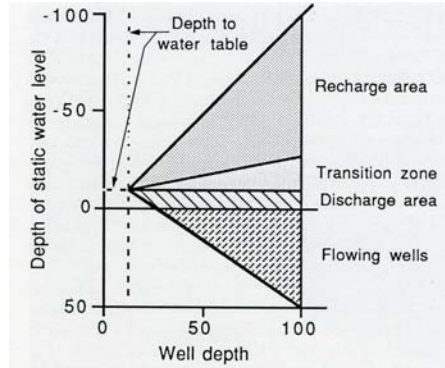
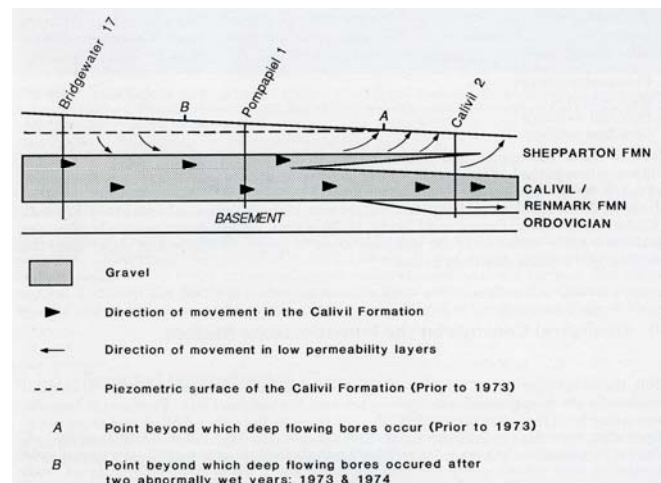


Figure 4.4 – Generalized plot of well depth versus depth to static water level (Freeze and Cherry, 1979)

Figure 4.5 – Flow pattern in the sediments of the south-central Loddon Plain (from Macumber, 1978a). 'Normal years' refers to the situation prior to 1973.



While aquifer wedging has a major influence on the regional discharge within the central Loddon Plain, an equally important factor, affecting both the Loddon Valley and Murray Valley systems in the lower Loddon Plain, is the major permeability decrease of the downbasin flow system as the gravel and coarse sand of the fluvial Calivil Formation ($K_x = 40\text{--}180\text{ m/d}$) are replaced by fine- to medium-grained sand of the marine Parilla Sand ($K_x = 1\text{--}2\text{ m/d}$) in the Kerang district. This major decrease in hydraulic conductivity virtually establishes a permeability barrier across the flow path of the regional Loddon-Murray system and is therefore seen as the most significant factor in the development of a groundwater discharge landscape in the Kerang-Swan Hill district.

Where vertical potentiometric gradients favour upward flow, the extent to which ground-water leaks from the Calivil Formation to the surface depends on characteristics of the overlying aquitard. These differ considerably on passing northward across the plains. At Bridgewater in the south, a zone of high permeabilities extends upward from the base of the aquifer to within 30 m of the surface, as shown on the gamma log (Figure 4.7a); this is in turn overlain by sandy units deposited in a fan head environment. In this region, the Calivil Formation is at best semiunconfined, with little impediment to vertical groundwater movement. On passing downbasin, there is a gradual decrease in the permeability of the aquitard, as the shoestring sands deposited as channel sediments give way to a far flood plain environment, with the resulting predominance of a clay lithofacies in the central Loddon Plain. Beyond Calivil, the Calivil Formation is overlain by a dense clay aquiclude about 80 m thick (Figure 4.7b), perhaps best developed in the Tragowel Plains area. Past Macorna, however, the Tertiary shoreline is crossed, and downstream from this point, the Calivil Formation is directly overlain by the marine Parilla Sand, itself an aquifer. Between Cohuna and Kerang, the Parilla Sand comes to within 15 m of the surface. As on the central and upper Loddon Plain, the standing water level of the Calivil Formation is either at or very close to the surface. Although the overlying units within the Parilla Sand are significantly less permeable than the basal units of the Parilla Sand, they would at the most be semiconfining and unlikely to provide much resistance to upward moving groundwater which has a positive hydrostatic head. This, too, is the case for the Quaternary units overlying the Parilla Sand. The movement of groundwater from deeper aquifers to the surface was documented by Poulton (1984), whose experiments in tile drainage in the Kerang area led him to conclude that "upward seepage of groundwater, from deep, semi-confined aquifer systems, was a major component in the total drainage water, particularly after the first three years of operation". He further commented that the results were "consistent with Macumber's proposition that, in the Loddon Plains, north of Macorna, the deeper aquifer systems, between the depths of 40 m and 90 m, are sources for groundwater discharge".

Between Cohuna and Kerang the Murray Plains are crisscrossed by a series of shoestring sands, associated with former courses of the ancestral Goulburn River (Macumber, 1968; Macumber and Thorne, 1975). The permeable soils of the Mead Ridge are developed on the levee systems of this stream. This situation is, therefore, similar to that found on the southern Loddon Plain, where shoestring sands have been shown to provide pathways for upward moving groundwater.

The variations in aquitard permeability, on passing down the Loddon Valley, are clearly seen in the gamma log section (Figure 4.7b). However, one significant difference is that, while groundwater discharge in the southern Loddon Plains is restricted to that provided by the Loddon Valley system, in the northern regions groundwater discharge is also fed from a Murray Valley system flowing from the east.



Figure 4.6a - Piezometer network in the Loddon Valley



Figure 4.6b - Potentiometric surface - Calivil Formation and Parilla Sand



Figure. 4.6c - Salinity distribution in the Calivil Formation - Loddon Valley Deep Lead

1	Bridgewater 15	2	Bridgewater 6
3	Pompaplel 1	4	Calivil 1
5	Mologa 1	6	Mincha West 12
7	Gunbower West 2	8	Gannawarra 4
9	Meran 10	10	Janiember East 18
11	Janiember East 16 & 17	13	Boort 24
12	Boort 25		

4.1.7 Characteristics of Groundwater Discharge Zones

The character of regional groundwater discharge areas was described by Meyboom (1962) for a regional flow system in south central Saskatchewan. The surficial evidence for groundwater discharge included saline soils and the associated phreatophytic and alkali tolerant vegetation, saline lakes and rivers, playas and salinas, springs and seepages. Later, Meyboom (1966a) correlated almost all the major areas of soil salinity in south central Saskatchewan with the discharge areas of regional groundwater flow systems.

Toth (1966), in a classification of field evidence for groundwater discharge from central Alberta, listed as well as above examples the observation of groundwater levels, flowing wells and the chemical quality of water. Freeze (1969b) noted the coincidence of high water tables and saline conditions in regional discharge areas of south Saskatchewan and suggested that special importance should be placed on salt tolerant phreatophytes as hydro-indicators. In many cases, evapotranspiration by phreatophytes is the sole form of groundwater discharge.

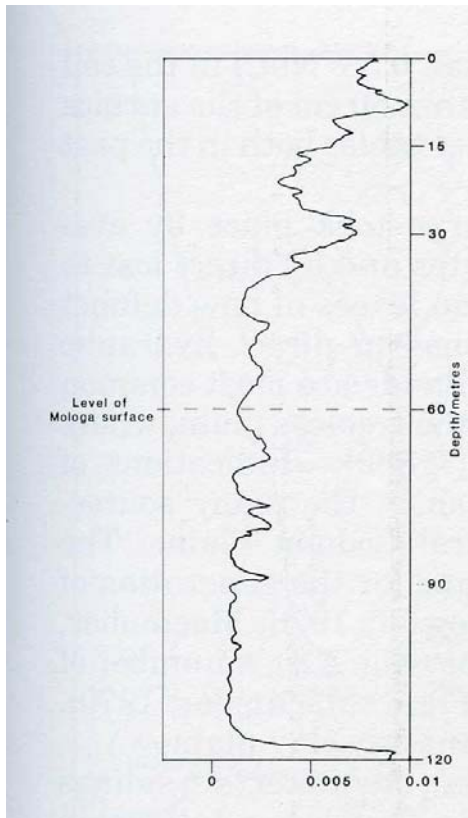


Figure 4.7a – Gamma log of Bridgewater 14

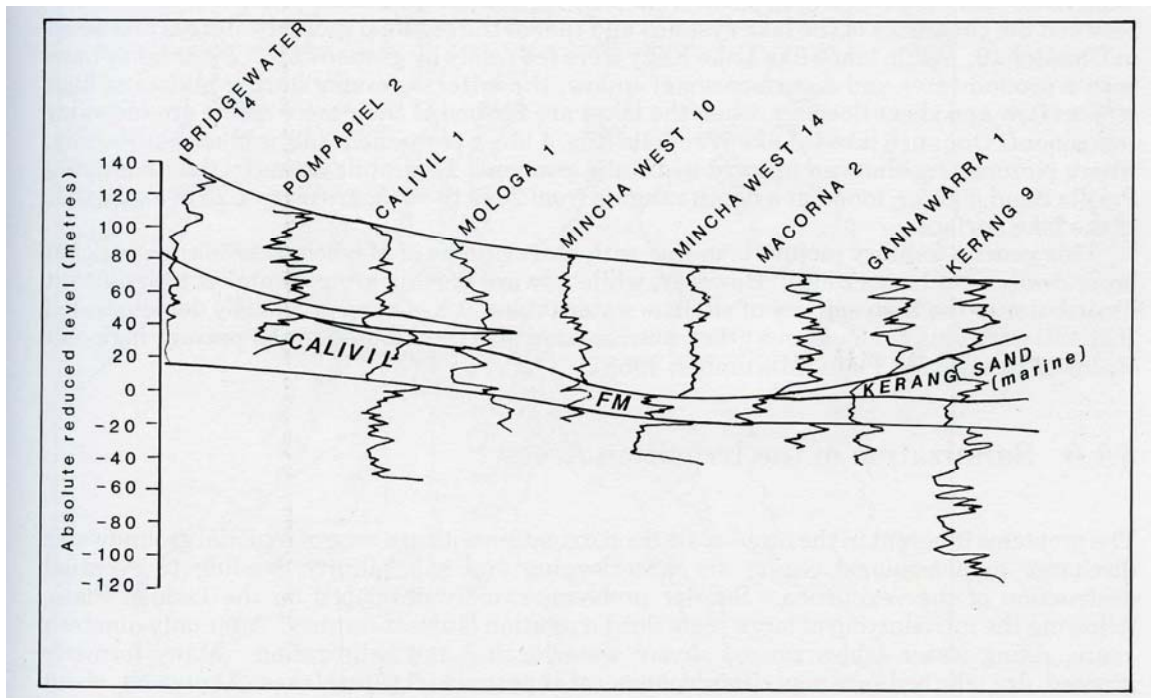


Figure 4.7b - Gamma log section between Bridgewater and Kerang.

Note how the aquitard to the Calivil Formation becomes increasingly less permeable on passing northwards across the Loddon Plain towards Minch West as the shoestring sand depositional environment gives way to a far flood plain environment. At Macorna, however, the higher energy Murray Valley sedimentary province is entered and the influence of shoestring sands on the aquitard is again seen, as is the incoming marine influence.

The position of the hinge line on the plains is of prime importance to agriculture, since, as it determines the recharge-discharge boundary, it marks the point beyond which areas occur where water which enters the ground cannot overcome the upward hydraulic gradient and will be subject to evapotranspiration in the vicinity of its point of entry. The Loddon Plain beyond Calivil, where the potentiometric surface of the Calivil Formation becomes artesian (used here to indicate flowing bores), shows many groundwater discharge phenomena; for instance, about one-third of the area in the mid-Loddon Plain has a salinity of greater than 0.3% NaCl in the near surface soils. Further north, most of the plains have greater than 0.5% NaCl in the soil (Skene, 1971). The prevalence of groundwater-deposited gypsum within 50 cm of the surface in soil profiles throughout the region reflects the presence of high water tables both in the past and at present.

Under presettlement conditions, regional groundwater discharge took place by evapotranspiration by scattered sclerophyll woodlands and phreatophytes and by direct loss to rivers and lakes. The woodlands (Skene, 1971) are best developed on levees of now defunct streams. Aquifers associated with the channel sands are at times in direct hydraulic connection with the more deeply buried regional aquifers. The woodlands are most common in the southern Loddon Plain but on passing northward give way to the treeless plain, where phreatophytes were present prior to European settlement (Curr, 1883). Indications of previous cycles of high water tables and saline conditions are seen in the many source-bordering clay dunes (lunettes), occurring in the lower and central Loddon Plain. The formation of lunettes requires high water tables and saline conditions for the generation of sand-sized clay aggregates, which are capable of being deflated (Bowler, 1976; Macumber, 1968). Apart from lunettes developed alongside ancient lake systems (Figure 4.8), a number of incipient lunettes have also been described from the lower Loddon Plain (Macumber, 1970). These formed adjacent to shallow saline depressions in the predominantly clay plain.

Effluent saline lakes are scattered over the Loddon Plain (Figure 4.9a), and in certain salinas (e.g., Lake Kelly), halite and gypsum were once regularly harvested. The close relationship between the chemistry of the lake systems and that of the regional groundwaters is discussed in Chapter 10. While lakes like Lake Kelly were fed solely by groundwater, other lakes have both a groundwater and a surface water inflow, the latter occurring during phases of high stream flow and sheet flooding, when the lakes are flushed of their more saline groundwater component. One such lake is Lake Wandella (Figure 4.9b), a permanent saline lake near Kerang, where piezometers show an upward hydraulic gradient. Hydraulic heads in the underlying Parilla Sand aquifer, found at a depth ranging from 20 m to 80 m, are up to 1.25 m above that of the lake surface.

This general salinity picture is in line with observations of Meyboom (1966a) on regional groundwater discharge zones. However, while upward moving groundwater is a significant contributor to the high salinity of shallow water tables, it has been previously demonstrated that salt additions from various other sources have also contributed to the present high-salt status of the Loddon Plain (Macumber, 1968).

4.1.8 Salinization of the Irrigation Areas

The problems inherent in the large-scale flood irrigation within a zone of regional groundwater discharge in a semiarid region are waterlogging and soil salinity, leading to eventual destruction of the vegetation. Similar problems rapidly developed on the Loddon Plain, following the introduction of large-scale flood irrigation late last century. After only nineteen years, rising water tables caused severe waterlogging and salinization. Many formerly grassed, dry lake beds are now either ephemeral or permanent saline lakes. At present, about one-third of the district is

severely salt affected (Figures 4.10 and 4.11) and virtually the whole of the central and lower Loddon Plain has a saline water table (15,000 - 50,000 mg/l) at less than 2 m. The immediate cause of this widespread salinization is clearly due to the increased water budget following the introduction of irrigation practice. While the heavy soils of the Kerang region may hinder the deep percolation of infiltrating surface waters, the upward hydraulic gradients in the groundwater systems counter downward percolation. This has led to a rapid rise in water tables, with the irrigation water being a major contributor to the overall water budget.

Therefore, the pre-irrigation situation on the Loddon Plain was one where groundwater discharge occurred in the northerly areas, with water seeping upward from the Calivil Formation into the Parilla Sand, and thence to the surface via the crisscrossing shoestring sand aquifers of the Shepparton Formation. The addition of irrigation water was to top up the groundwater system, with a resulting increase in the number and size of individual ground-water discharge points, leading ultimately to the development of a general high water table over the entire central- lower Loddon Plain.



Figure 4.9a – Lakes on the lower Loddon Plain. The pink lake on the left is saline and owes its color to the algae *Dunaliella salina*, whose cells accumulate carotenoids when grown in saline waters.



Figure 4.9b – Lake Wandella near Kerang. The piezometer Meran 10 on the lake shore has a static level of 1.2 m above lake level; it monitors the Parilla Sand aquifer.

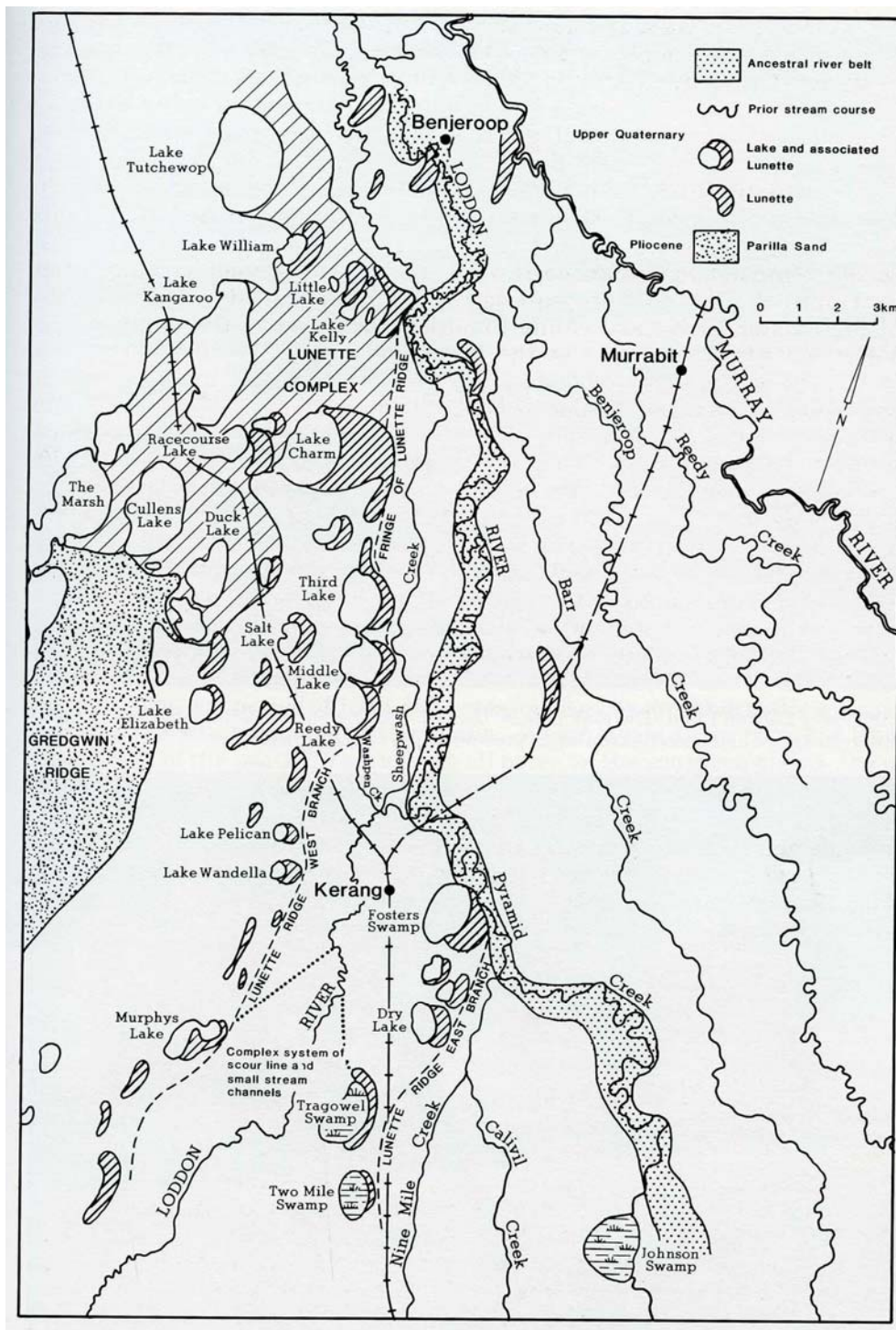


Figure 4.8 – Geomorphology of the lower Loddon Plain, showing the lake and lunette landscape. The Barr Creek is the stream with the low meander wavelength lying to the east of the Loddon River (from Macumber, 1968).

4.1.9 The Response of the Calivil Formation Aquifer to Short-Term Climatic Fluctuations

4.1.9.1 The impact of wet years on potentiometric head

Before 1973, a regular seasonal pattern of minor fluctuations in pressure levels was recorded from the Calivil Formation of the southern Loddon Plain, with a slight drop over the summer months and a similar rise in winter. The total amplitude of the fluctuation over a four-year period was only 0.6 m, peaking in late winter. During 1973-74, anomalously high rainfall was experienced in the catchments, and the plains were subjected to sheet flooding via the distributary systems of the Loddon River, Serpentine Creek and Bullock Creek (Macumber, 1978a).

In autumn 1973, the potentiometric surface of the Calivil Formation began to rise, and by early winter it had surpassed the highest levels reached during the four previous years; it continued to rise throughout the remainder of 1973 and peaked in late spring 1974. The rise was 3.5 m at Bridgewater, 3.7 m at Yarrayne, and 2.8 m at Pompapiel (Figure 4.12). As a result of the rising pressures, bores from Pompapiel to Calivil began to flow for the first time since observations were begun: the observation bore at Pompapiel rising to 1.10 m above the natural surface. The wet period of 1973-1974 was followed by an abnormally wet spring in 1975, which saw further sheet flooding of the plains. Water levels again rose, but from a base level well above that of the pre-1973 period. In all bores on the southern plains, the previous peaks had been surpassed by mid-October. The Pompapiel 1 standing water level had, by November, reached a point some 1.5 m above the ground surface. New levels at the Bridgewater 6 bore were only 3.7 m below the surface, a rise of 3.6 m over the levels of early 1973. Pressure levels began to decline in the drought of 1976, except in the Bridgewater 15 bore, where pressures peaked in November 1976. The Bridgewater 6 bore showed the largest rise in level of any bore on the Loddon Plain (4.0 m over the period March 1973 to November 1976), with lesser rises occurring in the Bridgewater 15 bore to the south (3.3 m) and the Pompapiel 1 bore to the north (3.1 m).



Figure 4.10 – Central Loddon Plain. Dead trees killed by saline water tables which lie within two metres of the surface.



Figure 4.11 – Salinization in the lower Loddon Plain.

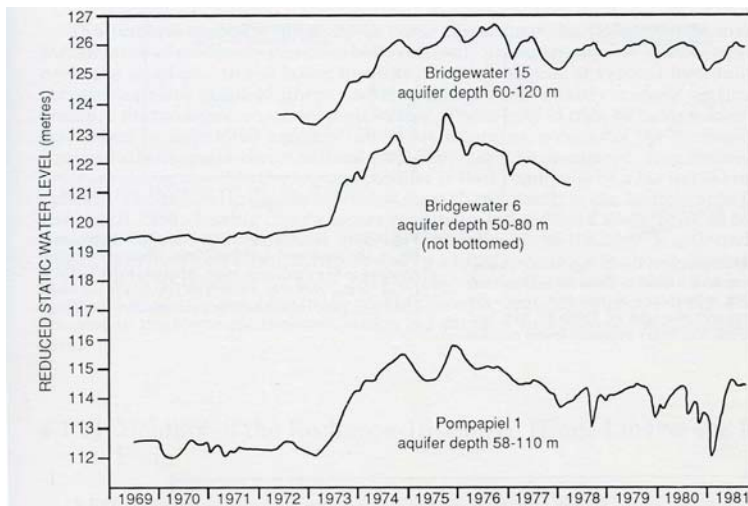


Figure 4.12 – Hydrographs of water levels of the Calivil Formation, southern Loddon Plain

On passing downbasin, the magnitude of the rise gradually declined, to be only about 0.2 m in the lower Loddon Plain, but again rising slightly, to be 0.5 m at South Cohuna on the Murray Valley system and at the Gannawarra 1 bore on the Kerang Research Farm. From Pompapiel onwards, all Calivil Formation bores were either flowing or very close to flowing.

4.1.9.2 The ineffectiveness of droughts in lowering regional groundwater pressures

In most bores, hydraulic heads peaked in 1975, with a slight subsequent gradual decline to reach new plateau levels significantly higher than their pre-1973 levels. This is clearly shown in the Bridgewater 15 piezometer on the southern Loddon Plain (Figure 4.12), where the peak levels in 1981 were 2.6 m higher than the equivalent readings of 1972. At Pompapiel, the seasonal peak levels were about 1.85 m above the previous peak, that is, about 0.4 m above the level of the surrounding plain, and even further above the level of the adjacent Pompapiel Creek bed, where groundwater outcrop was observed during the 1976 drought.

The 1983 drought, one of the most severe on record, had virtually no impact on the regional groundwater pressure levels, other than causing temporary falls in response to nearby groundwater pumping (clearly visible in the Pompapiel 1 and Bridgewater 15 piezometers - Figures 4.14 and 4.13). However, the drought was followed by a mildly wet spring in 1983, and groundwater pressures responded to recharge, reaching levels that surpassed the previous peaks of 1975 and establishing a new high plateau level, clearly visible in the Bridgewater 15 piezometer (Figure 4.13). Up to 1986, the levels did not fall from those of late 1983, and they now constitute a new plateau level from which future pressure level rises will be launched.

Similar patterns to those on the southern Loddon Plain were recorded in piezometers in the northern regions overlying the Murray Valley system, although hydraulic head rises were very much less than those recorded further south on the Loddon Valley system. For instance, in the Gunbower West piezometer, the peak level reached was in November 1974 (that in 1975 being only 0.02 m lower), a rise of 0.42 m over the peak of 1972. In 1979, the peak level was still 0.33 m above that of 1972, being about 1 m below plain level at this point (Figure 4.15). It is notable that the trend in the Gunbower West bore continues to rise, albeit very slowly. An explanation is that this piezometer lies close to the eastern fringe of the discharge zone and reflects rises occurring within the adjacent regional recharge areas.

It should be noted that on the southern Loddon Plain between Calivil and Bridgewater, peak levels are at times used to compare annual fluctuations, because the recent establishment of irrigation bores in the close proximity of a number of observation bores results in an exaggerated water level decline during the pumping season.

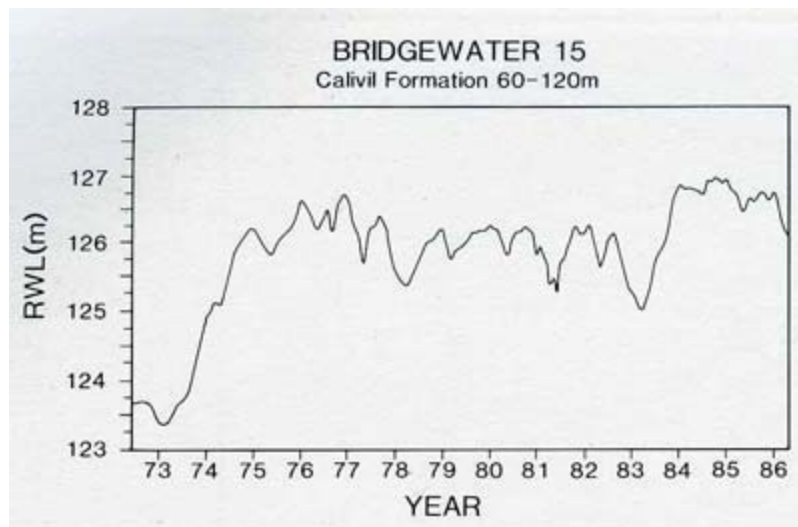


Figure 4.13 – Hydrograph of the Bridgewater 15 piezometer in the southern Loddon Plain.

Note the sharp rise in 1973-75 in response to the wet years. By contrast, the severe drought of 1982-83 has had virtually no effect, and today pressure levels remain high.

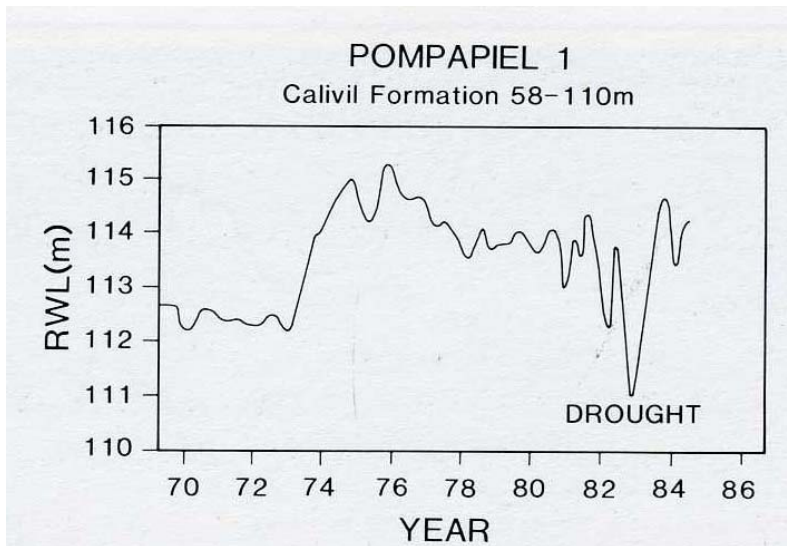


Figure 4.14 – Hydrograph of the Pompapiel 1 piezometer in the southern Loddon Plain.

This piezometer now lies in the discharge zone, whereas prior to 1972 it lay in the recharge zone. The low levels between 1980 and 1984 reflect nearby groundwater pumping.

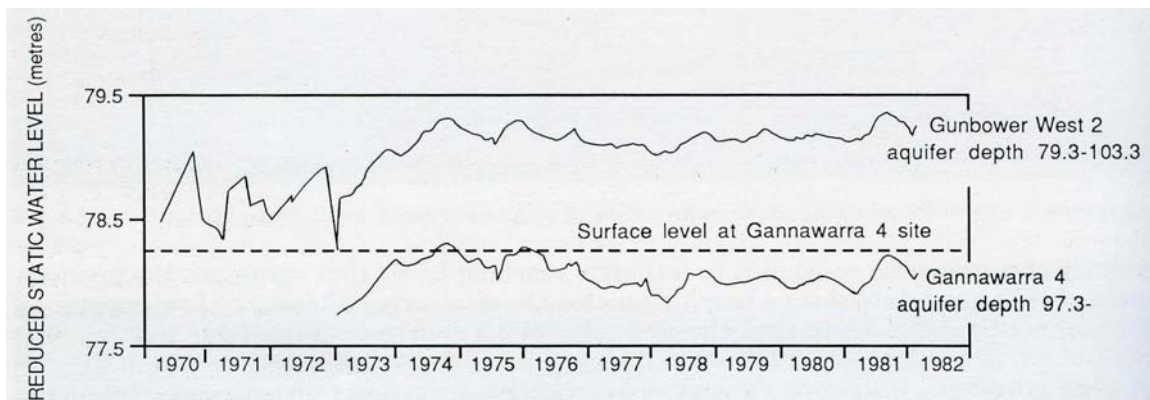


Figure 4.15 – Hydrographs of the Gunbower West 2 and Gannawarra 4 piezometers from 1970 to 1980, lower Loddon Plain

4.1.10 Hydrologic Equilibrium on the Loddon Plain

The suddenness of the rise, in 1973-75, in contrast to the slow subsequent rate of head decline, shows that natural groundwater recharge may be a rapid process, especially during periods of sheet flooding, as occurs on the Loddon Plain; the slow dissipation of pressures and the virtual nonimpact of severe drought show that natural groundwater discharge is a somewhat slower process. The pattern of overall pressure buildup in the Calivil Formation in the Loddon Valley is, therefore, either one of stepwise increments at times of exceptionally high surface water budgets towards new plateau levels, or a broad, sawtooth pattern where rapid rises are followed by very slow head decline, with the ultimate shape of the curve being determined by the periodicity of the recharge events.

Apart from the fluctuations in standing water levels in response to the more severe climatic aberrations, there is also a seasonal water level variation which is recorded in virtually all piezometers, whatever aquifer they monitor. As a consequence, all piezometers tend to fluctuate

in unison, despite their position on the plain. While their periods are essentially the same, their amplitudes vary, being dependent on their position on the plain and on their vertical position in the sequence (Figure 4.16). A seasonal pulsation occurs, with high levels being recorded between September and November, and low levels between February and April. The result is that the hydrographs for a bore such as Meran 10, monitoring an interval 46 m to 49 m in the Parilla Sand of the northern plain, is markedly similar to hydrographs of the Calivil 6 and Calivil 12 piezometers monitoring intervals of 66 m to 67 m and 31.9 m to 32.7 m respectively in the central Loddon Plain. This carries through to a point where even smaller out-of-season fluctuations also show, (seen in smaller peaks of April-May 1979).

The uniform response observed in bores throughout the Loddon Plain arises from the transference of pressure variations both vertically across the aquifer system and horizontally down the aquifers. In the latter instance, the high degree of vertical hydraulic connection between aquifers in the southern Loddon Plain permits a fairly uniform vertical transfer of pressure fluctuations, whatever their source. These may in turn be transmitted horizontally downbasin by individual aquifers into areas where, on account of the gradually increasing aquitard effectiveness, direct vertical responses might be dampened. Any vertical response to pressure changes within the regional aquifer is thus reinforced by a horizontal response from upbasin. The general influence of precipitation shows clearly in the hydrographs for the period 1977 until 1980. During this time, dry conditions existed until early 1978, to be followed by wet conditions for the remainder of 1978. The influence of the "dry" is reflected in the deep trough of 1978, which was in turn followed by a rapid recovery when most piezometers peaked in late 1978, at levels above their 1977 levels. While the extent of the 1978 low is perhaps increased by nearby pumping in the case of the three upper piezometers of Figure 4.16, this is not the case in the lower piezometers, which are situated well beyond the influence of irrigation bores.

4.1.11 Mobility of the Recharge - Discharge Hinge Line on the Loddon Plain

Prior to 1973, the bores in the Calivil Formation, in an area ranging from Calivil to beyond Kerang, were either flowing or very close to flowing. The hinge line between groundwater recharge and discharge was situated slightly to the south of Calivil (position A, Figure 4.5). However, with the water level rises of 1973 to 1975, the Calivil Formation became artesian from a point south of the Pompapiel 1 bore, where standing water levels were about 1.5 m above the surrounding plain. By late 1975, therefore, the hinge line could be placed slightly to the south of Pompapiel, about 20 km south of its previous position in the parish of Calivil (position B, Figure 4.5). This position is no longer in the irrigation district but in the dryland areas. At present (1986), groundwater pressures are at their highest since readings commenced in the mid-1960s, indicating that there has been no retreat of the hinge line, and instead, a slight upbasin advance has occurred.

The region between Calivil and Yarrayne must be regarded as an artesian transition zone, in which the hinge line migrates freely in response to pressure fluctuations in the Calivil Formation aquifer. The southern part of this transition zone is one where two major stream systems (the Loddon River and Bullock Creek) emerge from their highland tracts onto the plains. It is in this area where the Loddon fan (Macumber, 1968) begins, and where, in the early maps, Bullock Creek, which is well established further back within the highlands, is shown as petering out prior to rising again further downstream to become the Pyramid Creek. Under low-pressure conditions, the region is a recharge area, but under anomalously high pressures, as occurred after 1973, discharge conditions take over.

4.1.12 Groundwater Discharge on the Southern Loddon Plain

Upstream from Bridgewater the system is confined by basalt flows, and downstream at Calivil by clays; therefore, the higher hydraulic conductivity of the aquitard which is developed between Bridgewater and Calivil provides an escape valve for pressure buildup in the Calivil Formation.

The artesian pressures in the Calivil Formation are not restricted to the area of the lead trench,

but are transmitted widely across the southern plain by way of the sheet of coarse-grained sandy sediments, deposited over the Mologa Surface following the overtopping of the trench by Calivil Formation sediments (Chapter 2.5.5). The groundwater may then migrate upwards, along devious pathways of randomly occurring shoestring sands, previously deposited in a fan head environment close to the highland front. The direct distribution of pressure along the shoestring aquifers influences the ultimate pressure levels of shallow systems in the central lower plains. Since the shallow aquifers represent prior streams dying out on the plains, they have a decreasing carrying capacity by dint of lower downstream transmissivities and widths. High pressure levels in the deep Calivil Formation must, therefore, lead to greatly increased rates of water loss, from the shallow systems into the overlying clay layers. An indication of the groundwater loss from the Calivil Formation aquifer, occupying the Loddon trench between Bridgewater and Calivil, is readily obtained by use of a Darcy calculation, where

$$Q = KiA.$$

Q = rate of flow (m^3/d)

K = hydraulic conductivity (m/d)

i = hydraulic gradient

A = cross-sectional area through which flow occurs (m^2)

At Bridgewater, near the mouth of the Loddon Valley, the average transmissivity is $5.2 \times 10^3 \text{ m}^2/\text{d}$. At this point the Calivil Formation is 61 m thick, thus giving a hydraulic conductivity of 85 m/d . At Calivil, midway across the plains, the transmissivity is $895 \text{ m}^2/\text{d}$, and the aquifer thickness is 18.2 m, giving a hydraulic conductivity of 49 m/d (Table 4.1). Similarly, hydraulic gradients fall from 0.86 m/km , at Bridgewater, to 0.55 m/km , at Calivil.

Given that much the same aquifer porosity and trench width occurs at Calivil as at Bridgewater, the loss of water from the Calivil Formation can be calculated as a ratio:

$$\text{Flow (Calivil)} / \text{Flow (Bridgewater)} = 0.11$$

While this is only an approximation, it nevertheless provides an indication of the ground-water loss from the Calivil Formation in the region between Bridgewater and Calivil. This largely reflects the influence of the downbasin wedging of the Calivil Formation, which is not compensated for by any increase in hydraulic gradient or conductivity.

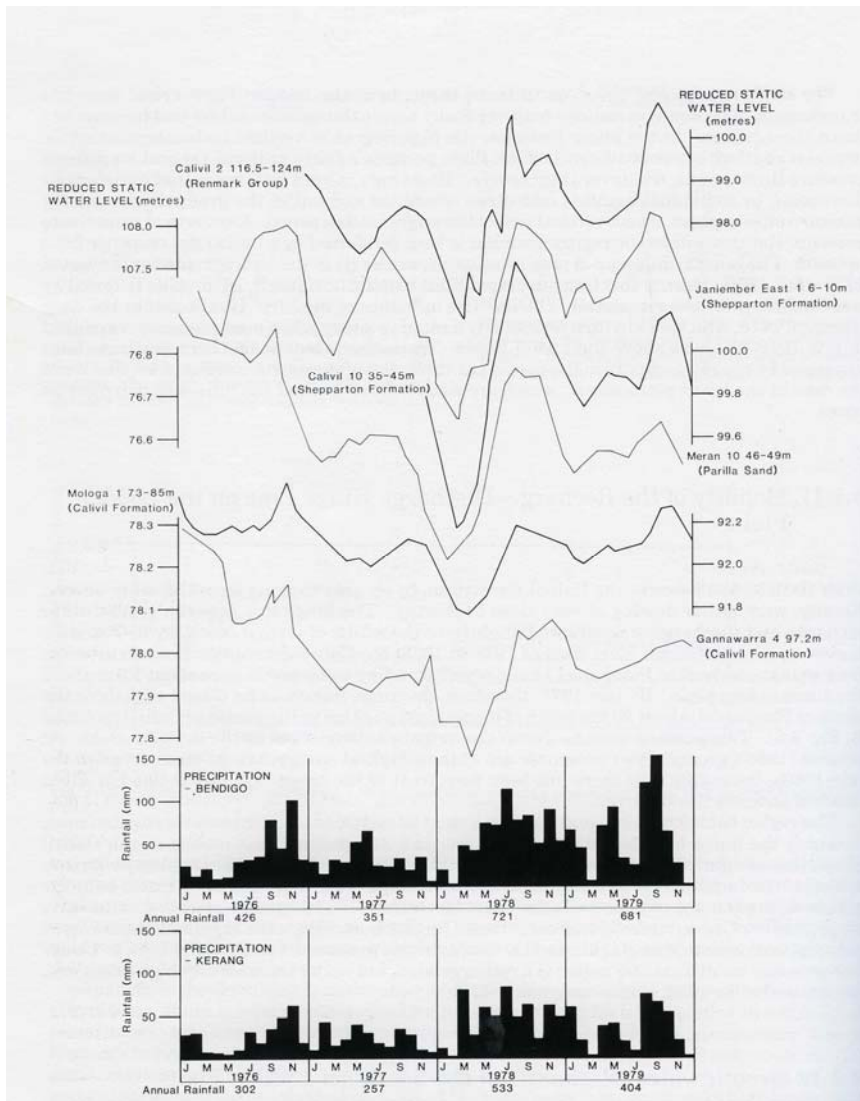


Figure 4.16 – Potentiometric head fluctuations for various aquifers beneath the Loddon Plain, with corresponding rainfall data

During phases of high pressure levels, as in 1973-75, the basal aquifer becomes artesian between Pompapiel and Calivil, with resultant upward water movement into the overlying shallow shoestring sands. The transmission of water along devious pathways of randomly crisscrossing shoestring sands spreads the area over which groundwater discharge occurs. Groundwater discharge to the surface would be particularly prevalent where the surface drainage systems intersect shallow shoestring sands. While it was not possible to separate groundwater outcrop from normal surface flow during the wet period, the following year, 1976, was one of drought, and groundwater discharge was observed in creeks rising on the Loddon Plain (Figure 4.14). This was the situation in the normally dry Bears Lagoon, which flowed saline groundwater throughout the 1976 drought (Figure 4.17). In the immediate vicinity of Bears Lagoon, shallow water tables appeared during the wet years in an unirrigated area of the plain that had previously been highly productive (Macumber, 1978a, 1978c). In the lowermost areas, groundwater outcrop was observed and these areas became saline (Figure 4.18). Bears Lagoon commenced flowing during this period and has since flowed intermittently.

A clear indication of the strong interaction between deeper and shallow aquifers was initially

obtained by using established private bores in the area. It was found that whenever the Janiember East 10001 flowing bore, situated about 1 km from the groundwater discharge zone at Bears Lagoon, was pumped (Figure 4.19), water levels rapidly dropped in a shallow piezometer (Janiember East 8010) in an adjacent discharge zone. Janiember East 10001 is screened over the interval 32.3 m to 36.3 m, while Janiember East 8010 monitors the interval 6 m to 10 m. Furthermore, analyses of aquifers occurring at varying depths throughout the district showed almost identical chemical characteristics to those of the shallow Janiember East 8010 bore (Macumber, 1978c).

4.1.13 Bears Lagoon - A Groundwater Discharge Feature of the Upper Loddon Plain

Bears Lagoon commences within a groundwater discharge zone (Figure 4.19) and clearly has its origin as a drain, operating at times of high pressure buildup in the aquifer system. In order to examine the relationship between surface flow in Bears Lagoon and hydraulic head in the deeper aquifers, a number of piezometers were established in deep, intermediate and shallow aquifers beneath the zone of high water tables adjacent to Bears Lagoon. At the same time, monthly chemical monitoring of each aquifer, and of the creek water, was also carried out.

The Janiember East 16 bore was drilled adjacent to the Janiember East 8010 bore and passed into Palaeozoic basement at 50 m. Gamma logs of the Janiember East 16 bore show that a 20 m thick basal sand overlies basement and comes to within 30 m of the surface, while other sand units occur at varying intervals between the top of the basal sand and the surface (Figure 4.20). The basal sand is the same unit developed in the Janiember East 10001 bore and represents the basal sand sheet which was deposited over the Mologa Surface following the overtopping of the lead trench by the Calivil Formation sediments (Chapter 2.5.5). It was traced westward from Bears Lagoon to a buried divide between the Loddon Valley and Powlett Plains leads, situated about 1 km away.

Piezometers were placed in the intervals 45 m to 32 m in the Janiember East 16 bore and in the interval 24 m to 19 m in a second bore, Janiember East 17, drilled alongside. An 8 m thick clay lens separates the zones tapped by the two piezometers. The two bores, together with the Janiember East 8010 piezometer (depth 6 m to 10 m), form a piezometer nest, which is regularly monitored. At the time of drilling, in September 1979, all three bores were flowing, with the deeper Janiember East 16 piezometer having a head about 1.8 m above the level of the adjacent saline area, where groundwater outcrop was present. The intermediate depth piezometer had a hydrostatic head 1 m above the nearby saline area and 0.9 m above that of the Janiember East 8010 piezometer. The hydrostatic heads of the three piezometers were monitored over a 30-month period up to 1982. During this time, the potentiometric level showed a seasonal fluctuation of about 1.8 m in all three bores. The pressure differentials between the three bores remained virtually constant, with the deepest piezometer, Janiember East 16, having the highest potentiometric head, which was always about 1.8 m above that of the shallow Janiember East 8010 piezometer. The intermediate piezometer, Janiember East 17, retained a head level intermediate between that of the deep and shallow piezometers (Figure 4.21). Throughout this period, the hydraulic head in the deep bore remained above the level of the adjacent, salt affected ground surface. The data indicates an ongoing upward hydraulic gradient, with groundwater flow from the deepest aquifer into the overlying aquifers. For much of the year, and especially during periods of peak hydraulic head, it leads to groundwater outcrop in the depressions and flow down Bears Lagoon. Chemical monitoring of the base flow of Bears Lagoon was carried out from 1975 up to the present, while analyses were also made of the Janiember East 8010 piezometer, and the Janiember East 16 and 17 piezometers after their completion. As in earlier analyses, where similar chemical character was observed in a number of bores in the vicinity, it was also the case with the Janiember 16/17 piezometer nest, which showed essentially the same water chemistry at all levels (Table 4.4). This situation continued over the two-year period for which observations were taken; it is interpreted as a single parent water feeding all aquifers, and in this respect supports the potentiometric evidence, which indicates

interconnection of all aquifers.



Figure 4.17 – Bears Lagoon, showing groundwater effluent. The flow commenced after the rise in groundwater pressure during the period 1973 to 1975. It has continued intermittently up to present.



Figure 4.18 – Salinized area adjacent to Bears Lagoon – the photograph was taken during the 1976 drought. The furthest piezometer is the Janiember 8001 piezometer.

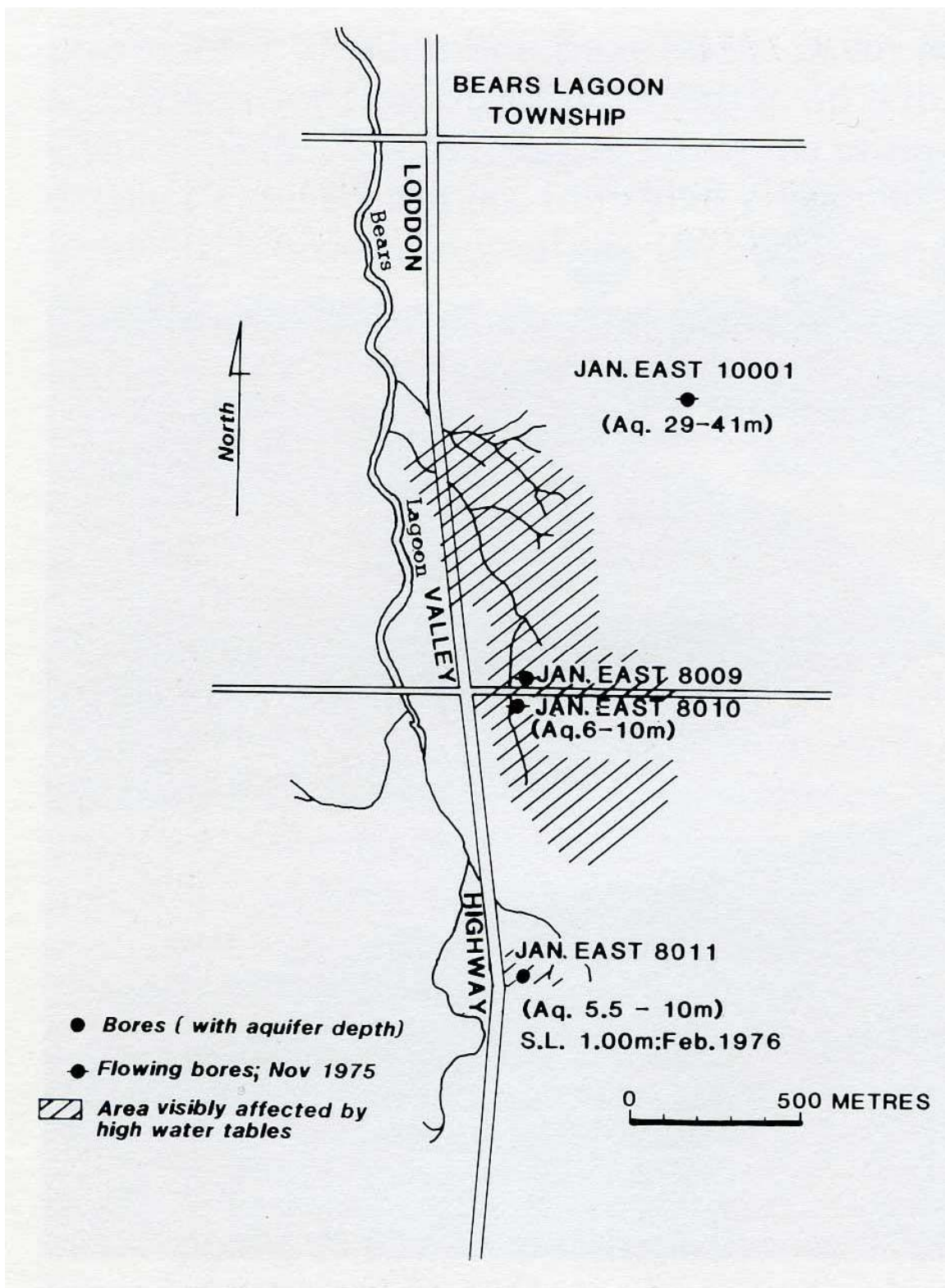


Figure 4.19 – Bears Lagoon, showing discharge area in November 1975

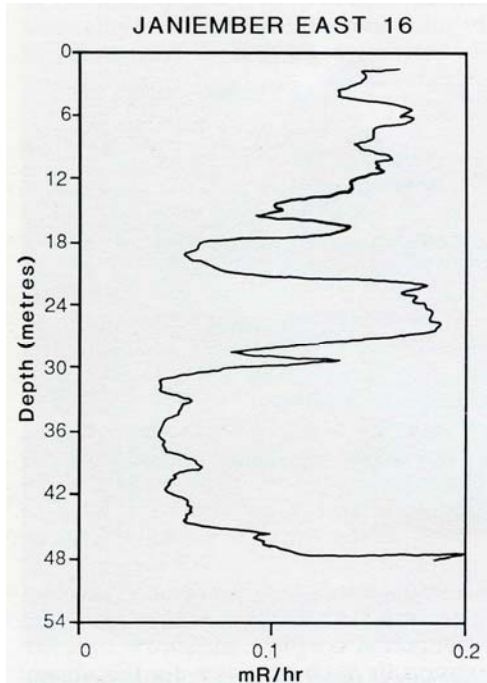


Figure 4.20 – Janiember East 16 gamma log

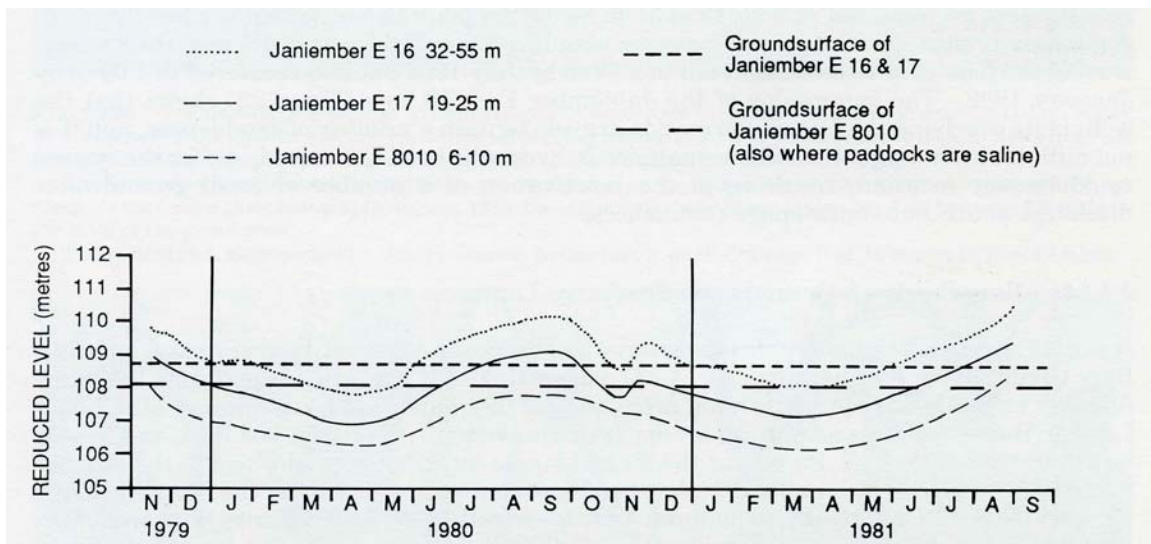


Figure 4.21- Hydrographs of the Janiember East 16, 17 and 8010 piezometer nest at Bears Lagoon. The hydrographs indicate a general upward directed hydraulic gradient, which is maintained throughout an annual cycle of pressure fluctuations.

Table 4.4 – Bears Lagoon piezometer nest analyses (13/1/81)

Bore No.	Janiember East 16	Janiember East 17	Janiember East 8010
Aquifer depth (m)	45 -32	25 – 19	10 – 6
Salinity (mg/l)	1557	1622	1641
Chloride	768	778	745
Carbonate	6	Nil	8
Bicarbonate	209	216	255
Sulphate	32	62	76
Nitrate	1	1	1
Calcium	16	25	25
Magnesium	79	77	85
Sodium	436	426	427
Potassium	7	35	7
Iron – soluble	<01	0.1	<0.1
pH	8.45	8.35	8.3

Throughout the monitoring period, higher salinity than that observed in the groundwater system was found in Bears Lagoon. In 1975, salinity was at its highest, being about 13,000 mg/l, compared with about 1,600 mg/l in the aquifers. This high salinity is seen as reflecting the initial outwashing of salts that had previously accumulated in the shallow, subsurface sediments and the soils as a consequence of capillarity. Once groundwater outseepage commenced, the accumulated salt was gradually flushed from the soils, with a subsequent decline in the salinity of the water in Bears Lagoon, which gradually leveled off at about 3,000 to 4,000 mg/l. This salinity is higher than the parent groundwater and suggests that about one-half to two-thirds of the water outseeping in the immediate vicinity of Bears Lagoon is lost by evapotranspiration. Bears Lagoon and its associated spring-heads represent a discharge point for the groundwater in the deeper aquifers; on air photos it is seen to have a distinct geomorphic character, which can be readily identified. The broad springhead zones of outseepage stand out clearly when seen from the air (Figure 4.22). The stream fed by the springs has a geomorphic character unlike other nearby streams, which form elements of the extensive distributary systems which have developed on past and present-day Loddon rivers.

The features which distinguish Bears Lagoon from normal streams on the plain are its tributary springhead feeders, its lack of natural levees, its absence of a coarse, bed-load channel component, and its low sinuosity. Its character is little more than that of a scour line across the plain. While the stream has been activated today as a consequence of the relatively recent rises in hydraulic head in the regional Calivil Formation aquifer, its origins are seen as occurring during extended periods of significantly higher water budgets, which existed in northern Victoria during the late Pleistocene (as recorded at Lake Tyrrell, see Chapter 3.3) and in the early Holocene (as recorded at Kow Swamp on the lower Loddon Plain, see Chapter 3.6). Under these conditions, much higher pressures would occur in the Calivil Formation aquifer, resulting in the upstream migration of the hinge line between regional groundwater recharge and groundwater discharge upstream to perhaps beyond Bridgewater, and with groundwater discharge occurring across the whole southern Loddon Plain. In support of this explanation, it can be observed on air photos of

the area that many similar springhead features occur in the upper Loddon Plain, as do many stream traces such as the Pompapiel Creek which has the same physiographic character as Bears Lagoon. During a parallel investigation of one such fossil spring system situated several kilometres west of Bears Lagoon, the Powlett Plains lead gravels were encountered at a depth of 55 m below the plain in the Janiember East 18 bore. A piezometer placed in the gravel showed a standing water level within 1.4 m of the surface. During the following 18 months, it fell to 2.32 m by July 1981 but had recovered to 1.09 m by January 1982. The gamma log of the Janiember East 18 bore (Figure 4.23) shows that the sediments overlying the Calivil Formation gravels contain a number of sand lenses, and it is not difficult to envisage that with a small rise in hydraulic head in the lead aquifer the system could become artesian, resulting in the reactivation of a number of fossil groundwater discharge points once outseepage commences.

4.1.14. Barr Creek - A Groundwater Discharge Landform

It is notable that the geomorphic characteristics observed on Bears Lagoon are also found on Barr Creek in the Kerang area (Figure 4.24), suggesting that this stream performs the same function as Bears Lagoon in draining areas subject to groundwater outseepage. Like Bears Lagoon, Barr Creek has no natural levees, it carries virtually no coarse bed load, and has an unusually low sinuosity. Its role as the largest single contributor of salt load to the Murray River from Victoria is well known (Gutteridge, Haskins and Davey, 1970). The Barr Creek salt load averages 170,000 tonne per annum, and this is reflected in the salinity increase which occurs in the Loddon River after its junction with Barr Creek just before it joins the Murray River (Table 4.5). Recent work by Gutteridge, Haskins and Davey et al (1985) shows that about 80% of the salt load in Barr Creek is of groundwater origin.

Barr Creek overlies a Murray Valley deep lead in the area where it is joined by the Loddon Valley system. In this region, the Calivil Formation aquifer is either artesian or virtually artesian. For instance, in the upper regions of Barr Creek, the water level in Zouza's bore (Gannawarra 8001, a private bore tapping the Calivil Formation) was about 0.6 m above ground surface. Gamma logs of the bore show that the Calivil Formation is overlain by Parilla Sand, which comes to within about 33 m of the surface. Further west, the Parilla Sand is within 15 m of the surface, and, under flowing artesian conditions, groundwater can readily migrate upward from the Calivil Formation into the Parilla Sand to emerge at low points on the surface, especially along incised stream channels and in deflated lake beds. The upflow of deeper groundwater into the shallow subsurface was observed by Poulton (1984) in his tile drain experiments at Kerang. The extent to which the present-day base flow in Barr Creek is derived from deeper groundwaters is not known, since, although conditions are conducive to this process, the effects are masked by other surface and groundwater additions due to the role of the creek as a regional drain.

The Janiember East 16 and 17 piezometers are established on the far side of the road, just to the right of the cross roads. At the time of this photograph (August 1979) the static water level in the Janiember East bore was 1.8 m above the level of the green area.

The springhead, represented by the green area, passes into a small drainage line, tributary to Bears Lagoon.



Figure 4.22a – Groundwater outseepage at Bears Lagoon.

Outseepage is clearly marked by green vegetation. Water in the dams is groundwater outcrop, the water table being almost at ground surface. Bears Lagoon flows from left to right, carrying only groundwater discharge, the salinity ranges from 3,000 to 4,000 mg/l.



Figure 4.22b – Tree covered lake floor of Middle Marsh on the Avoca River system, west of Kerang.

The forest died following the westward expansion of the saline discharge zone during the 1973-75 wet period.

TABLE 4.5 - Salinity along the Loddon River (18/8/65) (from Macumber, 1968)

Distance (km)	Locality	TDS (mg/l)	Cl ⁻	CO ₃ ²⁻	SO ₄ ²⁻	HCO ₃ ⁻
0	Newstead	173	46			66
21	Baringhup	246	70			66
43	Laanecoorie	391	150		22	79
56	Newbridge	432	176			85
72	Bridgewater	568	249		15	107
91	Serpentine	629	305		22	123
126	Durham Ox	842	411	6	23	136
145	South Canary Is.	650	304	7		109
157	Appin	723	352	4	16	127
187	South Kerang	920	396		15	138
222	Benjeroop+	1710	865	13	158	10

+ Downstream from junction with Barr Creek

4.1.14.1 Flow and Salinity Regimes in Barr Creek

A. Flow Regime

As a consequence of irrigation practice, there are two distinct flow regimes in Barr Creek - one occurring during the irrigation season, when the creek carries large volumes of irrigation drainage water, and the other occurring outside the irrigation season, centering on the winter period. In the irrigation season flows are generally high with peak flows reaching 500 - 600 ML/ week. During the winter months, stream flows are generally at their lowest, commonly stabilizing around 10-20 ML/week, when the creek has a large base flow component. Abnormally wet winters may result in high winter flows, with flow as high as 1100 ML/week (recorded during 1981). Flows in excess of 400 ML/week were recorded throughout the wet year, 1973.

B. Salt Regime

Salinity levels in Barr Creek inversely follow the flow rates, with highest salinities in winter, when the salinities commonly range between 6,000 and 18,000 mg/l (Figure 4.25). During the irrigation season, increased irrigation runoff in Barr Creek leads to fairly uniform salinities, ranging between 3,600 and 6,500 mg/l.

C. Salt Loads

Despite the very high winter salinities (at times of low stream flow, albeit high base flow), salt loads in Barr Creek closely echo flow rates and are normally at their highest during times of high flow, usually in the irrigation seasons (Figures 4.26 and 4.27). During wetter years, when winter flow is significantly higher, the bulk of the salt load passing down Barr Creek may be in winter. This was well illustrated during 1978/79, a wet but not abnormally wet period, when 170,000 tonne of the 250,000 tonne of salt passing down the creek did so in the nonirrigation season.

D. Character of Salt Flow Regime

While the bulk of the salt in Barr Creek is normally carried during the irrigation season, the distribution of the salt load is not even, but instead there is a large 'first flush' effect, with the largest volumes being carried during the initial phase of the irrigation season, when salt stored both on the land and in the streams as concentrated base flow is flushed into the system.

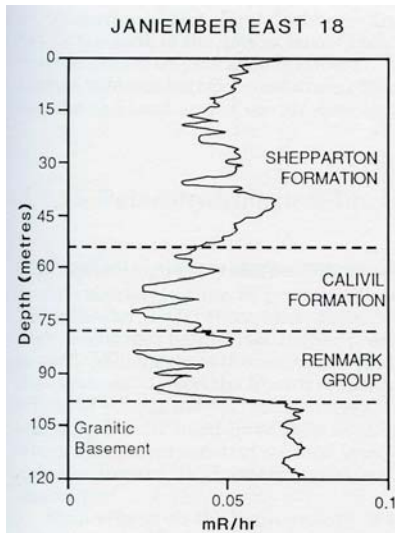


Figure 4.23 – Gamma log of Janiember East 18 bore

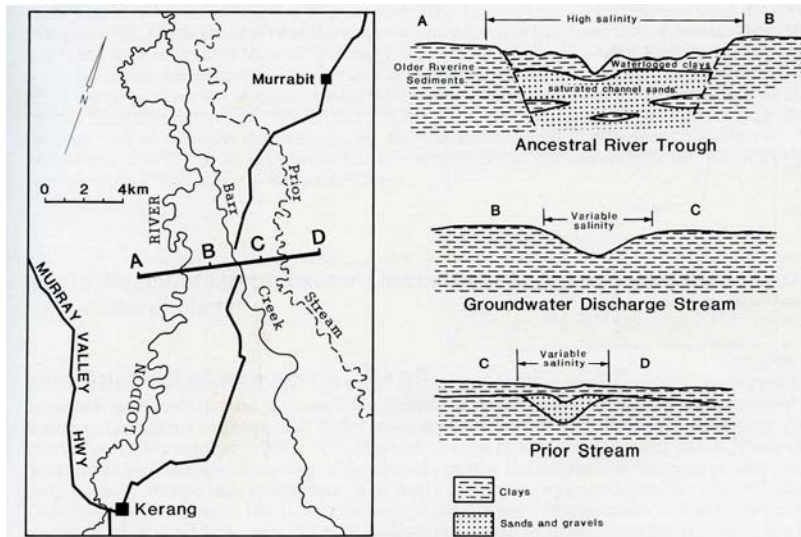


Figure 4.24 – This figure (from Macumber, 1968) shows the morphologic and sedimentologic characteristics of Barr Creek, which suggest that it has a similar origin and plays a similar role to Bears Lagoon, the groundwater fed stream on the upper Loddon Plain.

The low sinuosity of Barr Creek and the absence of any significant bed-load sediment are in sharp contrast to other stream types – the ancestral river and prior streams of the Loddon Plain.

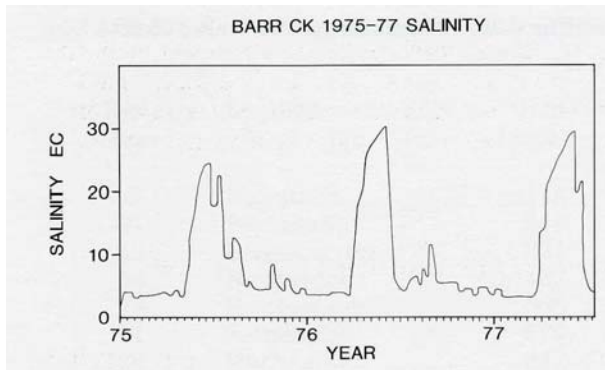


Figure 4.25 – Seasonal salinity variation in the Barr Creek

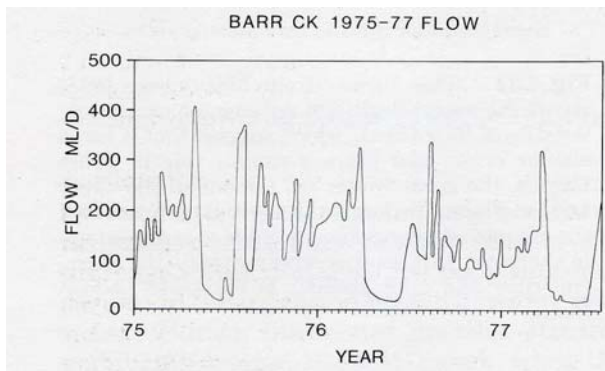


Figure 4.26 – Seasonal flow variation in the Barr Creek

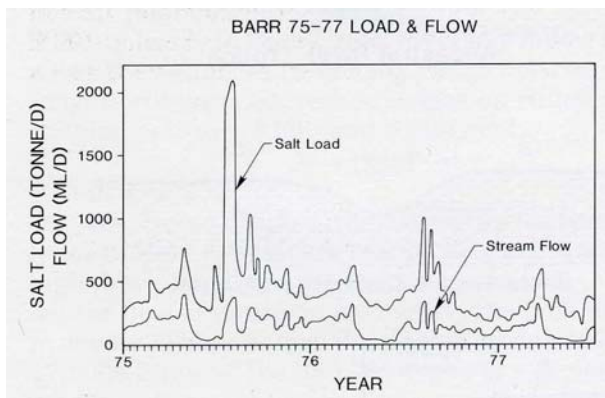


Figure 4.27 – Salt load versus flow in the Barr Creek

4.1.15 Palaeohydrological Implications of Unstable Flow Regimes

It is interesting to speculate on the effects of the late Quaternary periods of very much higher stream discharge than at present - these are well documented on the Riverine Plain (Pels, 1966; Bowler, 1967; Macumber, 1968; Schumm, 1968). There is some doubt as to whether the high discharges result from higher precipitation, or whether it is due to lower temperatures, or both. Whichever is the cause, it would lead to greater recharge of the regional groundwater flow systems. Given the known response of the aquifer to short-term climatic anomalies, the effects of similar events lasting several millennia or more would have been profound. Not only would hydraulic head have been considerably greater, but, more significantly, the point at which the potentiometric surface intersected the natural surface would have been pushed upbasin toward Bridgewater, making the entire Loddon Plain an area of groundwater discharge.

Such effects on the hydrogeology of the plains would have been dramatic. Groundwater outseepage would have been a significant feeder to already overloaded surface systems. This is clearly seen in the case of Bears Lagoon, whose origin is seen to date back to a time of general high-pressure levels and continuous groundwater discharge; it is almost certainly the origin of many of the small streamlet and drainage line traces now found crisscrossing the upper Loddon Plain. Under extreme conditions of waterlogging, the highly saline, shallow water table would have been at the surface in the central lower plains, with the resultant destruction of vegetation. (At present 90% of the shallow groundwater has a salinity of greater than 10,000 mg/l, and 40% is greater than 20,000 mg/l.) The lower central Loddon Plain would have been a virtual saline swamp, where water loss by evaporation led to ever increasing salinity levels. High water tables were in existence on the Loddon Plains as late as 8,000 years ago, as shown by the Kow Swamp study (Macumber, 1977). The reappearance of eucalypts along stream courses and in shallow depressions on the plains at about 8,000 years bp (Bowler 1968; Macumber, 1977) can be interpreted as a response to the decline in regional water tables on the Riverine Plain in early Holocene times.

4.1.16 Regional Groundwater Discharge - A Possible Factor in Stream Morphology

While the low meander streams and their springhead feeders are significant landforms on the Loddon Plain, perhaps the most striking feature of the river system is its transformation from a single meandrine course south of Serpentine to a leveed overbank flow distributary system to the north (Macumber, 1968). This marked change in river morphology is not just a recent occurrence but, instead, has been a feature of Loddon River systems throughout the Quaternary period. In the late Quaternary and early Holocene, a large delta-like alluvial fan, the Loddon fan (Macumber, 1968), was formed, commencing at Serpentine and extending 50 km across the Loddon Plain (Figure 2.23). A radiocarbon date from charcoal contained in the levee sediments of the fan (Figure 3.11) gave an age of 8,790 $\pm$ 120 (SUA-1320) years bp which corresponds with the period of high lake levels and high water tables recorded from Kow Swamp in the northern regions of the *Loddon* Plain (Macumber, 1977).

South of Serpentine, the Loddon River is incised below the level of the surrounding Riverine Plain, and is, at present, eroding into a terrace which shows the large wavelength meander scroll generally associated with the earlier ancestral river phases (Pels, 1966), and referred to by Lawrence (1975) as the Coonambidgal Formation. On approaching Serpentine, the terrace gradually rises in relation to the adjacent plain, until a point is reached at Serpentine where the terrace comes level with the plain. At this point, individual sandy point bars of the meander belt can still be distinguished; however, within a few hundred metres, the system rapidly changes into a system of leveed distributaries, the Loddon fan, which fans out across the plain. The present Loddon River and its principal distributary, Serpentine Creek, flow as lateral streams on either side of the fan, until a point is reached where they too fan out as distributary systems.

The change in stream morphology at Serpentine was previously attributed to changes in grade as the Loddon River passes from its highland tract to the plains (Macumber, 1968). It is now considered that the presence of a groundwater discharge zone across the path of the river may play a significant role in this process. This possibility arises from the observation that the point at which the river channel changes form coincides with the area where the underlying Calivil Formation first becomes artesian, and relic groundwater discharge traces appear on the surface.

The period when the Loddon fan was being aggraded coincides with the time when high water budgets were being experienced on the Loddon Plain (Macumber, 1977). Under these conditions, high hydraulic heads would have existed in the Calivil Formation, leading to considerable water loss from the surface and creating high water tables across the plain. In some respects, the hydrological regimes existing at the edge of the discharge zone are not too dissimilar to those existing in a subaerial deltaic environment, where a stream system passes into a lake or the sea, and this, then, provides the sedimentation base level to which the stream is graded. It is conjectured that a factor in the development of the Loddon fan was the presence of a base level of sedimentation in the upper Loddon Plain, imposed across the path of the Loddon River system by high water table conditions, induced by high hydraulic heads in the underlying Calivil Formation aquifer. The leveed distributary system on the Loddon Plain is, therefore, seen as having certain similarities to the leveed distributary systems developed in deltaic environments.

It is implied that, in certain instances, the depth of stream incision is influenced by the depth to the water table. The latter relationship is best seen wherever a pit is dug through relatively noncohesive sediments to a point below the water table. The extent to which this applies to stream channels depends, however, upon a number of factors, such as cohesiveness of the banks, erosive power of the stream, and its capacity to transport materials. A relationship between water table depth and depth of stream incision was shown by De Vries (1976) when he investigated stream networks developed in groundwater discharge areas of the Netherlands. In a groundwater discharge landscape developed in a fluvial setting, he noted that both stream spacing and channel depth decreased as the depth to the water table decreased. It is hypothesized, therefore, that the change in morphology of the Loddon River which led to the development of the Loddon fan was due to two factors:

1. Change in grade brought about by:

- a. the emergence from the highland tract onto the plain
- b. a base level of sedimentation, existing virtually at the plain surface, induced by the presence of a regional groundwater discharge zone across the path of the river

2. High water tables in the discharge zone which are conducive to shallow stream channels.

It is stressed that the above considerations on stream morphology are based on empirical observation and are difficult to prove to any absolute degree. They must, therefore, be regarded as tentative conclusions and remain at the level of hypotheses. It is notable that on the Campaspe Plain a late Quaternary fan system can be readily identified, having characteristics similar to the Loddon fan. This fan commences in the vicinity of the Rochester area, where it is predicted that, with the steadily rising pressure heads within the regional aquifer, there will emerge a regional groundwater discharge zone before the end of the century (Macumber, 1978c). On this basis, it is clear that the Campaspe fan commences in an area where regional groundwater discharge and high water tables existed in the past.

4.2 Campaspe, Goulburn and Murray Valleys

4.2.1 Hydrological Equilibrium in the Campaspe Valley

Eastward of the Loddon Plain is the Campaspe Plain. This is also underlain by a lead system containing Calivil Formation gravel, which forms the major aquifer beneath the plains (Figure 4.28). The Campaspe system differs from the Loddon system in that, while the drainage network within the highlands forms a tributary system, it is separated from the downbasin extension by a bedrock high (Macumber, 1978c). The main tributary to the system is the Huntly lead, which rises in the Bendigo area and passes eastward through the parish of Goornong to join a trunk lead system in the parish of Nolan. In the Campaspe Valley at Elmore, the potentiometric surface of the Calivil Formation is about 20 m below ground surface. Northward beyond Elmore, it rapidly rises with respect to the ground surface, to be within 5 m of the surface at Rochester, but again falls away to about 11 m in the parishes of Ballendella and South Echuca (Figure 4.28). Relative to the ground surface, therefore, a potentiometric high exists in the Rochester district (Macumber, 1978c).

On passing downstream (westward) beyond Echuca, the potentiometric surface again rises relative to the ground surface to become artesian in the vicinity of Gunbower. The recharge/discharge hinge line lies between Torrumbarry and Gunbower, the latter area lying within the regional discharge zone. By this stage the Campaspe system has merged with an east-west trending Murray Valley trunk system, which in turn is joined by a Loddon Valley system a little further to the west.

At present, shallow aquifer systems on the Campaspe Plain have a higher hydraulic head than the underlying deep systems, and groundwater moves downward into the deeper aquifers. While the pressure levels remain well below the ground surface, there are no problems arising from groundwater discharge from the Calivil Formation. However, a general rise in the potentiometric surface of the Calivil Formation aquifer of 0.1 to 0.3 m/yr has taken place over the past 75 years (Macumber, 1978c). The total rises have been in the order of 10 m to 20 m in this time.

Recent observations show that, in the parishes of Diggara, Rochester, Ballendella and Echuca South, rises over the period 1970-1981 in the Calivil Formation averaged about 0.2 m/yr, with the highest rates being 0.5 m/yr during 1973-75, when the potentiometric surface rose about 1.0 m to 1.25 m. In all instances, the hydraulic head continued to rise slowly until the 1983 drought, when intensive groundwater pumping helped stabilize the pressures. Since 1983, the number of groundwater pumps has greatly increased, and this has led to a quasi-equilibrium condition in the central Campaspe area.

The potentiometric surface approaches closest to the ground surface in the Rochester district, and with continued rise in pressures it would have become a discharge zone for the regional groundwater flow system within 40 years (Figure 4.29). In this respect, the Rochester district is comparable to the Pompapier district in the adjoining Loddon Valley, being especially sensitive to rises in regional groundwater pressures. While stability may have been achieved under the present pumping regime, any prediction on longer-term trends in the region must await the impact of the next very wet year.

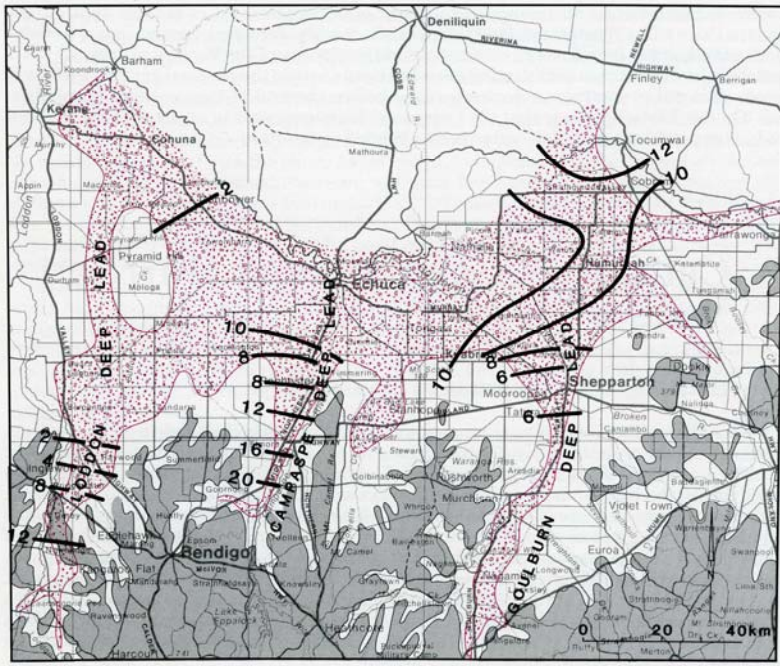


Figure 4.28a – Depth to the potentiometric surface of the Cavil Formation in Northern Victoria in November 1976 (from Macumber 1978c). More recent work has shown that the potentiometric surface in the vicinity of Rochester is now only 5 m below ground surface.



Figure 4.28b – Depth to the potentiometric surface of the combined Cavil Formation and Renmark Group aquifer, 1981.

4.2.2 The Goulburn Valley and Murray Valley Systems

There is some variation within the piezometers of the Goulburn Valley and Murray Valley systems; however, many piezometers show a general rise in pressure levels, ranging from about 0.03 m/yr up to 0.2 m/yr. Plots of a number of piezometer hydrographs, including linear regression analyses of the trends, are shown in Figures 4.30 to 4.35.

Within the upper reaches of the Goulburn Valley, the Tabilk 1 bore (Figure 4.30) showed an average rise of 0.15 m/yr from 1980 to 1985; however, data going back to 1969 shows that from 1969 to 1984 the pressure levels rose 2.94 m, giving an average rise during this period of 0.19 m/yr. Since the depth below ground surface to the potentiometric surface is only 6 m, at these rates regional groundwater discharge from the Calivil Formation aquifer will appear within 30 years. An expression of the discharge condition will appear sometime earlier, perhaps as increased stream salinity. Indeed, in the Tabilk 8007 bore, the pressure levels are already within 4.5 m of the surface.

While some bores within the central Goulburn Valley are fairly stable, as shown by the slow rise of about 0.03 m/yr in the Maroopna 6 piezometer, deep bores commonly show rises of 0.1 to 0.2 m/yr. The Waggarrandall bore (Figure 4.31) in the dryland area at the far east of the region shows a steady rise of 0.18 m/yr. Rises are well documented in the parishes of Tallygaroopna (Figure 4.32), Shepparton (Figure 4.33 - note that the deeper piezometer is in the bedrock aquifer), Katunga, Balkamaugh, Waaia, Maroopna and Barwo. As is also the case in the Murray Province to the north, the potentiometric surface is in most cases within 10 m of the ground surface. Given the present rate of rise, it can be predicted that virtually the entire Goulburn Valley will become a regional discharge zone within 70 to 90 years. This estimate is based on a constant rise of 0.15 to 0.1 m/yr, and does not take into account such major recharge events as occurred between 1973 and 1975.

Across the Murray Valley province in the northern parts of the plain, the rise in deep groundwater pressures is consistently about 0.2-0.1 m/yr, as shown in the Strathmerton and Narioka bores (Figs 4.34 and 4.35). In these areas, the potentiometric surface is 9 m to 11 m below ground surface (Figure 4.28), suggesting that the area will have artesian pressures within 50 to 100 years. As in the case of the central Goulburn Valley, this does not take into account any major recharge event.

4.3 Hydrological Equilibria in Riverine Settings

4.3.1 Factors Influencing Hydrological Equilibrium

Climatic factors such as increased rainfall over the last 20 years or so have been suggested as the cause of the regional groundwater pressure rises. However, despite minor climatic fluctuations since 1900, the rises in the Campaspe Valley have been fairly uniform during different time periods - 1900 to 1958, 1925 to 1974, and since 1975. It seems that, while wet periods may have a significant impact upon pressure levels, in the long term these effects average out. Indeed, if pressure rises as have occurred since 1900 were projected back several hundred years, it would have required an initial condition in which the aquifers were dry. This proposition is untenable and leads to the inescapable conclusion that pressure rises are a relatively recent phenomenon. That is, the steady rise in the potentiometric surface of the Calivil Formation aquifer has required an increased rate of aquifer recharge since European settlement. This increase, although not great in terms of absolute water budget, is sufficient to tilt a delicately balanced equilibrium towards a position of steadily increasing groundwater storage.

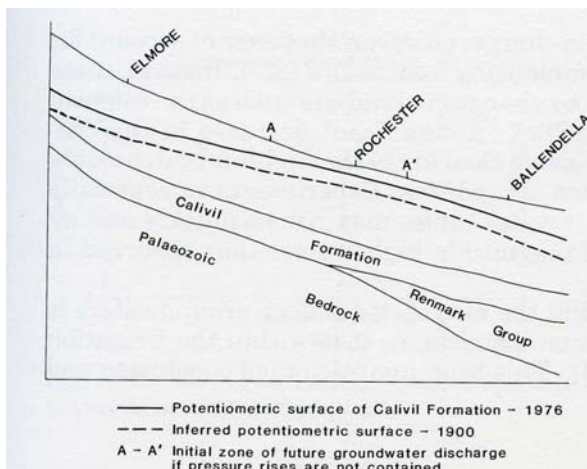


Figure 4.29 – Diagrammatic section showing the rise in the potentiometric surface of the Calivil Formation aquifer in the Campaspe Valley since 1900, and the predicted future zone of regional groundwater discharge if groundwater levels are not controlled.

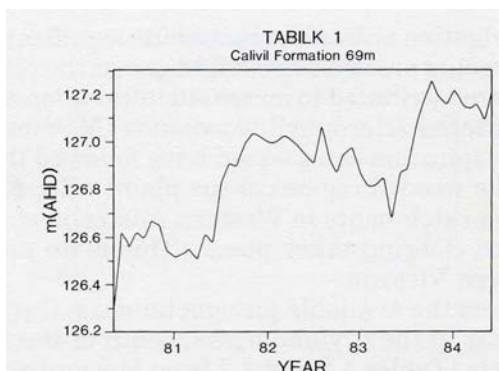


Figure 4.30 – Hydrograph of Tabilk 1 piezometer, southern Goulburn Valley

Given that there was no significant increase in precipitation immediately prior to 1900 when the standing water levels along the Campaspe lead system were first recorded, it follows that the increased recharge to the aquifer was due to changes in land and water usage. Increased water

accessions brought about by irrigation practice are readily detected in the irrigation districts; however, outside the irrigation districts, in the dryland areas, the potentiometric surface of the Calivil Formation is also steadily rising. Since the plain tract of the Campaspe lead is tectonically separated from the highland tract (Chapter 4.2.1), it is clear that virtually all recharge to the Calivil Formation occurs on the plain.

TABLE 4.6 - Pressure rises in the dryland province

Position	Rise (m/yr)	Period	Comment
Tabilk 1	0.15	1980-84	Goulburn Valley
Tabilk 1	0.19	1969-84	
Goornong	0.24	1900-58	Campaspe Valley
Nolan/Elmore	0.20	1926-75	Campaspe Valley
Diggara/Elmore	0.26	1925-75	Campaspe Valley
Elmore	0.20	1900-83	(From Tickell, 1977)
Bridgewater 15	0.14	1972-84	Loddon Valley
	(rose 4 m 1973-75)		
Pompapiel 1	0.08	1969-84	Loddon Valley
	(rose 3 m 1975-84, now in discharge zone)		
Waggarandall 1	0.18	1975-84	Goulburn Valley

Irrigation accessions contribute significantly to recharge; however, they cannot account for the ongoing pressure rises right across the plain, commencing from before 1900. Instead, these rises are attributed to increased infiltration caused by changes in land use such as the removal of scattered sclerophyll woodlands (Macumber, 1978c). A significant decrease in the evapotranspiration rates would have followed the extensive clearing of the highland catchments and the wooded regions of the plains after European occupation. Experiments on especially cleared catchments in Western Australia show that water tables may rise as quickly as 2 m/yr once clearing takes place. This is an order of magnitude higher than that observed in northern Victoria.

From the available piezometer data, it seems that the rate of rise of deep groundwaters is as great in the dryland areas, south of the irrigation districts, as it is within the irrigation districts (Tables 4.6 and 4.7 from Macumber, 1984). This is an unanticipated conclusion and an area where further work is warranted.

TABLE 4.7 - Pressure rises in irrigation districts

Position	Rise (m/yr)	Period	Comment
Tallygaroopna 10	0.15	1975-84	Goulburn Valley
Narioka 2	0.11	1975-84	Murray Valley
Strathmerton 62	0.17	1975-84	Murray Valley
Campaspe Plain	0.13	1970-84	average of 38 bores
Ballendella 5	rose 1.08 m over 1972-76		
Ballendella 8	rose 0.8 m over 1972-76		
Echuca South 4	rose 1.8 m over 1973-76		
Rochester (average)	rose 1.5 m over 1973-76		

4.3.2 Recharge Areas and Recharge Events

Tickell and Humphrys (1986) suggest that recharge into the regional aquifers may range from 5 mm/yr up to 40 mm/yr in the Campaspe Valley; Williamson (1983) suggested that it maybe as high as 100 mm. The records of rising regional groundwater pressures suggest that recharge occurs right across the plain - indeed probably occurring in virtually all areas where discharge is not present (and at times even in these areas). Yet, major recharge events are clearly recorded in the short time during which detailed observations have been taken.

These observations have led to the conclusion that, while infiltration may be a fairly universal process, there are nevertheless broad zones of preferential recharge (and discharge) where throughflow is significantly higher. On the Riverine Plain this arises from the presence of crisscrossing and interconnected shoestring sands. The zones of preferential recharge occur at a macroscopic and microscopic level: for instance, the fan head environment where streams emerge from the highland front is an example of the former; points of virtual aquifer continuity from the deep systems to the surface (as at Bears Lagoon in the Loddon Valley), are examples of the latter.

Zones of preferential recharge are important for groundwater accessions, but perhaps even more important are the discrete recharge events as occurred throughout northern Victoria in 1973-75 (and probably in 1954 and 1966). At such times, widespread sheet flooding on the plains leads to rapid groundwater recharge. It is unlikely that during these events there is any natural process that can prevent massive recharge. Apart from general recharge across the plains, the channel deposits of streams are also a significant part of the recharge system, having an extended role under high flow conditions, when the silts and other finer materials normally blanketing the coarser bed load, are taken into suspension, leaving the coarse channel sediments exposed in the base of the stream, and hence increasing outflow through the streambed.

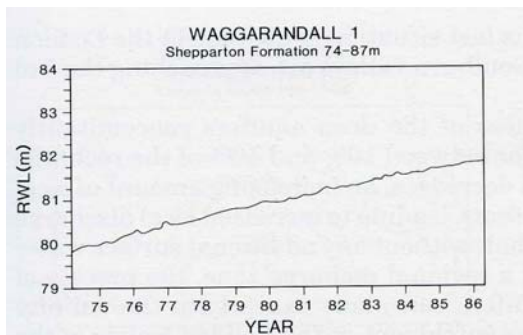


Figure 4.31 – Hydrograph of Waggarandall 1 piezometer, northeastern Goulburn Valley

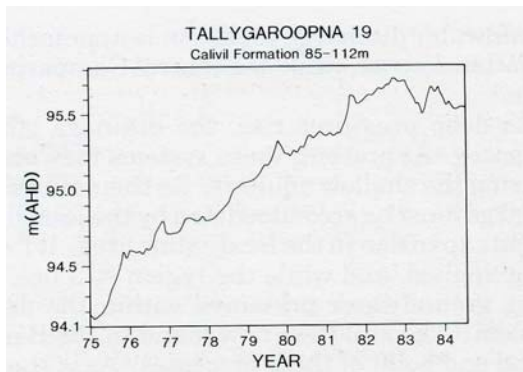


Figure 4.32 – Hydrograph of Tallygaroopna 19, Goulburn Valley

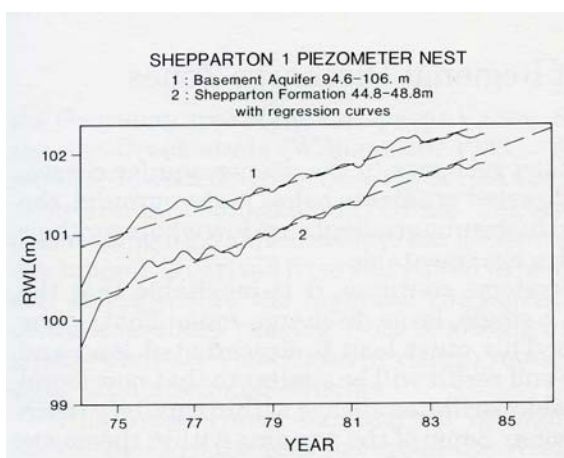


Figure 4.33 – Hydrographs of Shepparton 1 piezometer nest

4.3.3 The Influence of Rising Groundwater Pressures on Surface Processes

It is clear that recharge on the plains influences pressure levels in the deep regional aquifers; yet, the reverse is also true, and here the deeper pressures influence the recharge rates, and, therefore, the extent to which deep drainage provides a means of naturally disposing of excess water entering the shallow groundwater regime. For instance, given the continued gradual rise in potentiometric head, the rates of recharge to the deep aquifers must also steadily decline, as vertical hydraulic gradients decline in response to pressure rises within the deep aquifers. That is, the vertical gradient when the potentiometric surface is, say, 20 m below ground surface, would be about 0.25, falling to 0.06 when at 5 m. It also follows that there is a comparable decrease in recharge to the deeper aquifer. Once the potentiometric surface approaches to within 5 m of the surface, the rate of deep recharge will halve with only a further 2.5 m rise, and so on. Beyond this point, vertical gradients fall off very rapidly as a regional groundwater discharge condition is approached. This last situation is the case in the Loddon irrigation district, while the central Campaspe and Goulburn valleys are approaching the 5 m stage.

As deep pressures rise, the drainage effectiveness of the deep aquifers concomitantly decreases. At present, these systems may account for between 10% and 40% of the recharge entering the shallow aquifers. As their effectiveness decreases, an increasing amount of local recharge must be accommodated by the local flow systems, leading to increased local discharge and an expansion in the local saline area. It follows that, without any additional surface water being applied, and while the region still lies within a regional recharge zone, the process of rising groundwater pressures within the deep aquifers can alone exacerbate the salinity problem. The situation now found in the Barr Creek catchment of the Loddon Valley is the logical end point of these processes. As is shown in the Gutteridge, Haskins and Davey et al. (1985) report, it is clear that the bulk of the salt entering Barr Creek is from groundwater.

4.3.4 The Evolution and Expansion of Regional Discharge Zones

Once a regional discharge zone comes into being net recharge to the deeper aquifer ceases. While there is still some potential for downward directed gradients below higher ground, the net groundwater movement is upward, and vertical hydraulic gradients are low whatever their direction. If not already present, high water tables are inevitable.

If the long-term trends in the groundwater systems continue, it is inevitable that the Riverine Plain in Victoria will eventually become a single, large discharge zone. That is, the Tertiary/Quaternary aquifer system will be full. This must lead to exacerbated land and stream salinity. In the case of irrigation areas the end result will be similar to that now found on the Loddon Plain, where a stable saline water table oscillates always within capillary reach of the surface, but rising and falling with the seasons. Some of the streams within the newly established discharge zone will take on a role similar to that of Barr Creek and Bears Lagoon, which are essentially effluent groundwater drains.

On the basis of available data, regional groundwater discharge zones will first appear in the central Goulburn Valley, and the Campaspe Valley if deep groundwater pumping fails to suppress the pressure rises. These conditions will gradually extend throughout the irrigation districts, with the regional hinge line at Gunbower extending eastward to meet an expanding discharge zone from further upbasin. On the basis of the Loddon Valley model, it appears that while it may take some time for a regional discharge zone to become established in a drainage basin, once present, then its expansion is almost inevitable under the present high water budget regimes. The situation now found on the Loddon Plain - a steady state, saline environment undergoing lateral expansions at its perimeter, especially during wet years - is taken as being the end point of this process.

4.4 Dryland Areas the Plains and the

Highlands

The potentiometric data from the southern plains and the adjacent highland valleys suggests that regional discharge conditions may readily appear in the dryland areas upbasin from the irrigation districts. This is the case at Bears Lagoon in the Loddon Valley now and will be the case in the Tabilk area of the Goulburn Valley within 30 years; it may already be the case in the Goornong area of the Campaspe Valley. Some indication of the effects can be obtained from the Axe Creek study (Williamson, 1983). Williamson demonstrated that the volume of salt passing down Axe Creek over the period 1966-1981 averaged 15,000 tonne/annum. This is 9% of the annual salt load of Barr Creek. Yet, Axe Creek is only one of a number of similar streams discharging high salt loads into the major rivers of northern Victoria. The salt source in these catchments is derived from salt stored in bedrock aquifers underlying the catchments. Indeed, salt stores within the highland catchments are relatively high with salinity levels of the bedrock aquifers ranging from 3,000 mg/l up to 30,000 mg/l TDS. Salt stored within the aquifers has been mobilized by rising groundwaters and brought to the surface where it affects soils, vegetation and streams.

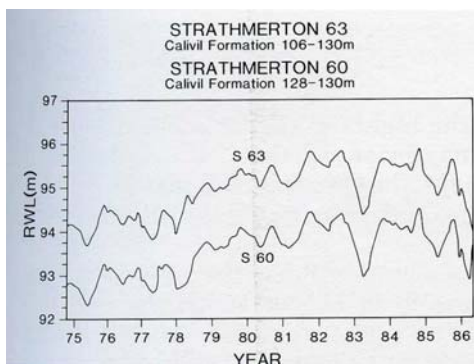


Figure 4.34 – Hydrographs of Strathmerton 63 and 60 piezometers, Murray Valley

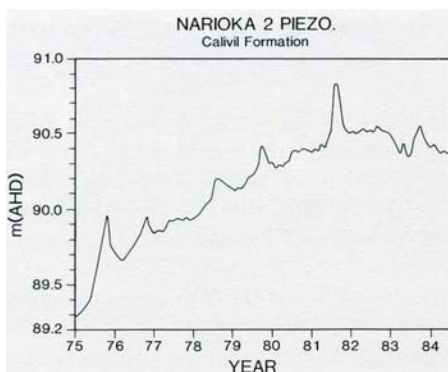


Figure 4.35 – Hydrograph of Narioka 2 piezometer in the Murray Valley

This process may be observed in highland areas across northern Victoria, from the Strathbogies in the east to beyond the Wimmera River (Figure 4.36) in the west, including the Grampians and the Dundas Tableland (Figure 4.37). The salinity problem is twofold in the highlands: firstly there is land salinization, caused by rising water tables, and secondly there is stream salinization. Tributaries of the north flowing Goulburn, Campaspe, Loddon, Avoca, Wimmera, Richardson and Avon rivers are affected. In the southwestern parts of the state, many streams are already brackish or saline. It is no coincidence that virtually all southward flowing streams in western Victoria are brackish or saline, as they are further along the track and represent the natural end point towards which the northern Victorian tributaries are evolving.

On the basis of the form of the groundwater flow system, the highland region may be subdivided into three distinct salinity provinces (Table 4.8, Figure 4.44), based on hydrogeologic and geomorphic criteria (Macumber and Fitzpatrick, 1987).

1. Province MA : This province, making up from 10% to 30% of the highlands, is developed on local groundwater flow systems.
2. Province MB : This province is developed on regional groundwater systems developed within the fractured rock aquifers of the highland region.
3. Province HC : This province covers the major river valleys within the highlands. Unlike the other highland provinces, it is developed on unconsolidated sediments infilling the highland valleys. In general, groundwater flow systems are regional. An example of regional salinization in province HC is in the central Avoca Valley along Bet Bet Creek where the potentiometric surface of the Calivil Formation is virtually at the ground surface, creating a regional discharge zone within the highlands and causing land salinization. In this area, the Avoca River has a significant base flow component and is often brackish, with salinities reaching 5,000 mg/l.

4.4.1 The Saline Cycle of Erosion

The rise in water tables in the bedrock aquifers across the highland region is leading to the salinization of many small and large catchments of the major river systems. It is widespread and occurring in such diverse landscapes as the Strathbogies, the Grampians (Victoria Valley), the Dundas Tableland, the Heathcote-Colbinabbin Range, and, more generally, across virtually the entire central highlands.

The initial effects of catchment salinization are the loss of vegetation in the valley bottoms, the dispersion of clay 'B' horizons leading to the breakdown of soil structure, and the commencement of soil and gully erosion. Incision of the gully creates a natural drain, facilitating outflow from the groundwater systems and dropping the water table to below the capillary fringe, thus relieving the effects of adjacent land salinization. It concentrates the outflow and in doing so, enhances stream erosion and valley development. Groundwater discharge is controlled by geomorphic and structural/lithologic factors and, therefore, is not restricted to valley bottoms, but instead commonly occurs at other distinct points in the landscape. One well-recognized outflow area is at the break of slope on the valley sides (Dyson, 1983).

The effects of these processes within the highlands are the reverse of those described on the Loddon Plain, where the high regional water tables cause modification of stream morphology by acting as a base level of erosion, inhibiting the depth of stream channels, and so enhancing the development of braided or anastomosing systems. Instead, in the highland settings, the steeper ground surface gradients and the resultant higher energy of the streams enable more rapid removal of sediment, permitting gullying to proceed below the water table, which in turn is locally lowered in response to increased outflow.

The link between high water tables, salinization and gully erosion is clearly visible throughout the southern Australian landscape, and this process may perhaps be the most important single factor in the modification of highland landscapes during Quaternary times. As a consequence, the saline cycle of erosion is seen as a significant process, especially in semiarid landscapes with similar hydrogeologic and geomorphologic characteristics to those occurring across the Australian continent.

4.5 Extensions to the Mallee Region of Northwestern Victoria

The increased infiltration in the highland valleys and on the plains is also occurring in the Mallee areas to the west of the Riverine Plains, where even more extensive clearing of the land has taken place. Unlike the Riverine Plain, the Mallee region of northwestern Victoria with its aeolian landscape is underlain by the 60 m thick unconfined Parilla Sand aquifer. Groundwater within the Parilla Sand is generally saline and commonly ranges from about 25,000 to 50,000 mg/l. In the northern Mallee the water table is close to the surface, and groundwater often outcrops as small salinas scattered throughout the landscape (Macumber, 1980a). Although originally covered with a low eucalypt woodland, this has largely been removed and the land is now extensively used for cropping.

Over the last 50 years many lower depressions within the undulating landscape have become salinized (Figures 4.38 and 4.39), suggesting a gradual rise in water tables. Rowan (1971) noted an expansion in the size and number of salt pans over the period 1946 to 1963, which he attributed to rising groundwaters. Although no long-term potentiometric data is available, the effects of rising water tables can be seen in the number of previously dry, timbered depressions and former lake beds, which are now virtual salinas or gypsum flats developed over ground-water sinks. The trees scattered across the depression are now either dead or dying, attesting to an ongoing change in the hydrological regime. Wherever these neosalinas occur, they are found to occupy the flat floors of more ancient salinas, formed during an earlier cycle of high water tables (Figure 4.40).

Other effects produced by the rising water tables include the cutting of roads built across lower depressions (Figure 4.41), thereby forcing their realignment around the rim of the depression or, at times, their abandonment; and the salinization of stock dams. The latter are particularly susceptible and may be converted into minisalinas. One farm dam in the Nangiloc-Colignan area to the east of the Raak boinka (see Figures 4.42 and 4.43), had a salinity of 180,000 mg/l, when measured in December 1978. Elsewhere, where dams become salinas, the surrounding land may often be converted into gypsum flats.

As in the highland region, the Mallee can be divided into regional (MB) and local (MA) groundwater flow systems. However, in this case, over 95% of the region is underlain by a saline regional flow system, for which there is no apparent solution at present to rising water tables.



Figure 4.36 – Dryland salting in the Dundas Tableland



Figure 4.37 – Dryland salting in the Wimmera River catchment



Figure 4.38 – Salinized area between Manangatang and Ouyen. The recent rise in water tables has killed the vegetation and salt now effloresces on the surface. The regional groundwaters have a salinity in excess of 30,000 mg/l. This area became saline following the 1973-75 wet phase.



Figure 4.39 – Similar landscape to Figure 4.38 but showing a more advanced stage of salinization, with saline groundwater outcrop



Figure 4.40 – Salinized area at Bonoener (northeast of Raak boinka).

This saline flat lies within a subdued linear dunefield and was formed during past cycles of groundwater discharge. The dam and dead trees attest to the earlier non-saline status of the area. Under the present hydrologic regime the dam is now virtually a neosalina.



Figure 4.41 – Salinized area near Chillingollah, east of Lake Timboram.

The road shown crossing the depression is now no longer passable and has been abandoned. Groundwater outcrops in the background, and seepage occurs at points well above this level.



Figure 4.42 – Salinized stock dam at Bonoener.

The dam became saline during the 1973-74 period, and when sampled in December 1978 it had a salinity of 180,000 mg/l



Figure 4.43 – Salt efflorescing on higher ground near Bonoener.

Included in this area are stock yard which became unusable. In the distance a fully salinized depression can be seen.

4.6 The Salinity Problem - An Overview

4.6.1 The Problem

Throughout southern Australia there is a delicate balance between groundwater and surface water systems, each of which makes up an aspect of a single two component downbasin flow system. The balance is such that even a small increase in surface water budgets may cause a significant rise in groundwater pressures. The rising groundwaters can, in turn, affect surface processes, leading to a rapid shift in environmental equilibrium, which manifests itself in widespread salinization.

We today experiencing a wide-ranging and fundamental readjustment in hydrological equilibrium, brought about by the impact of European technology on what was previously an untouched landscape. A major factor in this shift has been the extensive clearing of trees from both the highlands and the plains, together with the introduction of large-scale irrigation. This has led to vastly increased runoff from the catchments, resulting in accelerated erosion and sedimentation in differing parts of the many drainage basins. However, the most drastic changes stem from increased infiltration which has led to gradual filling up of the aquifer systems. The rising groundwaters are bringing salts stored in the aquifers to the surface, where they are causing salinization of soils, vegetation and streams.

This process is occurring throughout the landscape, in virtually every regional setting, including the Riverine Plain of northern Victoria, the Wimmera and basalt plains of the western districts, the highlands stretching from the Strathbogies in the east to the Grampians in the west, the Mallee, and most irrigation districts. Indeed, the high salinity of streams throughout southwestern Victoria is a consequence of groundwater outseepage, and this same process is now having a major impact on tributary systems feeding the major rivers of northern Victoria.

In irrigation districts, the increased groundwater accessions are greatly exacerbated by imported water, and even more rapid water table rises occur. In addition, salt is imported in irrigation water and gradually accumulates in the soils and shallow groundwaters. The dilemma, central to the salinity problem in irrigation districts, is the choice (or balance) between external salt disposal and on-farm techniques (which imply a 'living with salt' option). However, on-farm options whilst they are important control techniques cannot prevent salt accumulation. Furthermore, they can at best limit the rate of water table rise; they cannot stop it short of capillary reach of the surface. Instead, subsurface drainage (e.g., groundwater pumps and tile drains) and salt export are essential means of preventing the eventual establishment of permanent high saline water tables.

The annual cost of salinity to Australia in lost productivity is estimated to be about \$95 million; while the cost to Victoria alone is about \$50 million, which could rise to about \$100 million in 50 years if nothing is done.

Rises in regional groundwater pressures are best documented in the Riverine Plain, where groundwater monitoring has been taking place since the 1960s. For instance, in the Goulburn/Murray plains, rises varying from 0.1 m/yr to 0.2 m/yr are recorded from a number of piezometers, from both dryland and irrigated regions alike. Data from the Campaspe Plain show that the average rate of rise of regional pressures since 1900 is about 0.25 m/yr. The deep regional aquifers are important subsurface drains; however, as groundwater pressures rise, the effectiveness of these deep drains decreases, and there is a consequent spread of local groundwater discharge and, hence, an expansion in the salinized areas. If these pressure rises were to continue at present rates, it could be predicted that the entire Riverine Plain in Victoria - dryland and irrigation areas alike - will become a zone of regional groundwater discharge within 100 years. At present, this situation occurs only in the Kerang irrigation district, where saline water tables are everywhere within capillary reach of the surface.

Similar rates of groundwater pressure rise have been recorded from dryland highland areas, while CSIRO experiments in Western Australia, on newly cleared forest, show that rates of water table rise may at times be as great as 2 m/yr. On the basis of anticipated rising groundwater trends in the Mallee region, Allison (CSIRO, Adelaide) has predicted a significant increase in salinity pickup by the Murray River; however, there is still much conjecture about the timing and extent of this prediction.

The rises in groundwater pressures are continuing unabated throughout much of the landscape, with little evidence that equilibrium is near being reached. In those instances where equilibrium is close, it is an essentially brackish to saline equilibrium, which has in the case of the plains (e.g. the Kerang Region) a high, increasingly saline water table, which oscillates seasonally, but lies always within capillary reach of the ground surface. In undulating settings, as in the Mallee and the highlands, the inevitable result of continuing water table rises is saline depressions, or brackish to saline land and stream systems respectively. That is, under the present land use techniques, the trends in rising water tables will continue, with the gradual expansion in land and stream salinization being inevitable and essentially irreversible. The salinity problem is seen, therefore, as being fundamentally a groundwater problem. The control of water tables is essential if salinity is to be avoided.

The present trends are not unique, but represent the most recent of a number of similar saline cycles, which have occurred naturally in the past in response to fluctuations in climate. These ancient saline cycles have left their imprint in the soils, sediments, landforms and groundwaters. For instance, the extensive lake systems and associated landforms of both northern and southwestern Victoria are partially a legacy of earlier saline cycles. These earlier saline events provide a basis not only for a better understanding of the process of groundwater – surface water interactions now being experienced, but may also, perhaps, provide a means of predicting the direction, extent and form of the present man-induced saline cycle.

Indeed, while the general trends in land and stream salinization are very clear, the eventual extent and degree of salinization is still being assessed. The ultimate effects will obviously depend on the extent to which remedial action can be taken. However, it is clear that, while much can be done in some areas, there are some important areas where little or nothing can be done, and in others it will take considerable time and effort to stop, let alone reverse, the present trends.

4.6.2 Salinity Provinces

There are a large number of options available for salinity control; however, particular options are only technically feasible in areas with specific physical characteristics. Accordingly, Victoria has been divided into 15 salinity provinces, based on hydrogeologic and geomorphologic criteria (Macumber and Fitzpatrick, 1987). Each province has its own small range of technically feasible control options. Additional salinity provinces may be defined in the future, depending upon the need. Distinct provinces exist in the Mallee, the highlands, the plains, southwestern Victoria and the irrigation districts. The 15 salinity provinces are described in Table 4.8, and their locations are given in Figure 4.44.

There are fundamental differences in the approaches to salinity control in the irrigation and the dryland areas, and the basic division between dryland provinces and irrigation provinces is retained. The dryland areas are further subdivided on the basis of the nature of the groundwater flow system into: local systems, or type 'A', where recharge and discharge areas are close by, and regional flow systems, or type B', where recharge is some significant distance from the discharge area. Over 70% of the dryland area has regional groundwater flow systems.

TABLE 4.8 - Victorian salinity provinces

A. Dryland provinces			
Province		Characteristics	
Mallee			
Mallee	MA	Perched dune seepage, associated with local flow systems	
Mallee	MB	Regional saline water table, e.g., Raak Plain Highland	
Highland	HA	Local flow systems in basement aquifers, e.g., Bourkes Flat	
Highland	HB	Regional flow system in basement aquifers	
Highland	HC	Predominantly regional flow system in valley-fill alluvium, e.g., Avoca/Bet Bet valley	
Plains			
Riverine	PR	Predominantly regional flow systems	
Basalt	PB	Regional and some local systems, especially in stony rise areas	
Wimmera	PW	Mainly regional systems, e.g., Lake Buloke	
Coastal	PC	Coastal plain, mainly regional, but also local systems	
Dundas			
Dundas	DA	Dundas Tableland, essentially local flow systems	
B. Irrigation provinces			
Province		Groundwater Salinities*	
		Shallow Aquifers	Deep Aquifers
Loddon province	IL	Saline	Saline
Campaspe province	IC	Brackish	Fresh
Goulburn province	IG	Brackish	Brackish
Murray province	IM	Fresh	Fresh
Sunraysia province	IS	Saline	Saline

Irrigation provinces are divided along catchment lines, but also differ in the character of their underlying groundwater bodies. For instance, the saline, shallow and deep groundwaters of the Loddon province (IL) preclude groundwater pumping with reuse. By contrast, the fresh, deep and shallow aquifers of the Murray province (IM) enable conjunctive groundwater and surface water usage, and do not pose the same immediate off farm disposal problems as in IL, or even the Goulburn province (IG).

Whilst there are a wide range of salinity control options available for a number of differing saline settings, it must be clearly understood that control techniques successful in one salinity province will not necessarily work in a different province. Failure to appreciate this point is largely responsible for confusion within the community about the application of salinity control techniques. For instance, techniques for salinity control in a local flow system within a highland province HA will not have the same effect on a regional highland flow system HB. They could be largely ineffective if carried over into an irrigation province, and a total waste of effort and resources if used to control water tables in a Mallee regional province MB. On the other hand, they may be effective on the perched water tables associated with local flow systems developed within Mallee dunes (province MA). Fresh <1,000 mg/l, brackish 1,000-4,000 mg/l, saline >4,000 mg/l.

4.6.3 Salinity Control Options

Salinity control options may be divided into four broad categories:

- (i) Options designed to reduce groundwater accessions
- (ii) Options providing subsurface drainage to lower water tables
- (iii) Options designed to facilitate the adaptation of agriculture to high water tables - living with salt option
- (iv) In some circumstances it is clear that there is no suitable control option. In these cases, the best course of action may be to do nothing.

4.6.3.1 Control options for reduction of accessions

Options aimed at reducing groundwater accessions address the fundamental cause of the salinity problem.

Methods for reducing groundwater accessions include:

- reforestation
- establishment of deep-rooted pasture
- introduction of minimum tillage and direct drilling
- landforming
- improved irrigation systems and techniques
- improved irrigation scheduling
- improved surface drainage

A. Dryland area

Dryland salinity provinces include the Mallee region, the highlands, the plains (basalt, Riverine and Wimmera) and the Dundas Tableland.

In the case of the local flow systems, where cause and effect lie in close juxtaposition, treatment of the cause may have a fairly rapid response in the adjacent salt affected areas. This is the case at Bourkes Flat in the Avoca catchment, where the successful use of lucerne has reversed the previously upward-directed water table trends.

In regional flow systems where treatment must be on a regional multicatchment scale, salinity control may take many generations to achieve, and in some instances it may be virtually unattainable. This is the case for the regional saline aquifers of the Mallee. No control methods short of extensive replanting are available: even with a total cover there can be no reversal of the present salinity pattern.

In dryland regions, the reduction of accessions to the groundwater is likely to be the most appropriate salinity control technique for most areas. The most economically viable options are agronomic/ agroforestry and on-farm measures, aimed at reducing access. Groundwater pumping is not a viable option for most of the dryland provinces, where in general brackish to saline groundwaters and low permeability aquifers predominate. Exceptions to this occur in the

cases of the highland valley province (HC) and certain regions of the Riverine Plain province (PR), where suitable aquifers permit localized irrigation.

B. Irrigation districts

The irrigation salinity provinces correspond to the major irrigation districts: Loddon Valley (IL), Campaspe Valley (IC), Murray Valley (IM), Goulburn Valley (IG) and Sunraysia (IS). Each province has a different combination of hydrogeological characteristics which influence control options (Table 4.8). The shallow aquifers refer to the Shepparton Formation, while the deep aquifer refers to the Calivil Formation and, at times, the Renmark Group. In irrigation districts, it is likely that methods to reduce accessions will, at best, slow the rate of rise of the water table until a discharge equilibrium is reached. they cannot, per se, prevent water rises short of capillary reach of the surface.

4.6.3.2 Subsurface Drainage

If rises in water tables and the subsequent waterlogging and salinity problems are to be prevented, subsurface drainage is essential. The aim of subsurface drainage is to lower water tables to below 1 - 2 m from the surface (depending on soil and crop type), either by groundwater pumping or tile drains. Successful groundwater pumping requires suitable aquifers and, in cases where pumped groundwater is used for irrigation, salinities of less than about 3,000 mg/l.

It is likely that less than 35% of all irrigated land in the Riverine Plain will have suitable aquifers for groundwater pumping. Subsurface drainage provides the best and probably the only means of preventing saline agriculture in irrigation districts. However, in the long term at least, the greatest constraint on salinity control by subsurface drainage is the problem of disposal of groundwater effluent. High priority must therefore be given to the evaluation of different disposal options.

It follows that:

- subsurface drainage is essential in irrigation areas if the eventual rise of water tables to within capillary reach of the ground surface is to be prevented
- subsurface drainage and reuse without salt export from the district leads to gradually increased soil and groundwater salinities
- disposal of saline water is the key issue in salinity control by subsurface drainage.

4.6.3.3 On-farm techniques - living with salt options

It is likely that, in the absence of subsurface drainage, and given the continued present irrigation practices, saline agriculture will become a default option in many areas. At present, this situation exists throughout the Kerang region, where on-farm techniques are the basis for any long-term strategy. Under saline agricultural regimes farm productivity is lower, adding extra pressures on farm viability.

Where subsurface drainage is not practical, the living with salt option aims at optimizing farm productivity under high water table conditions which must inevitably arise in irrigated areas not having subsurface drainage. These techniques, aimed at optimizing farm productivity, are equally applicable in nonsaline regimes.

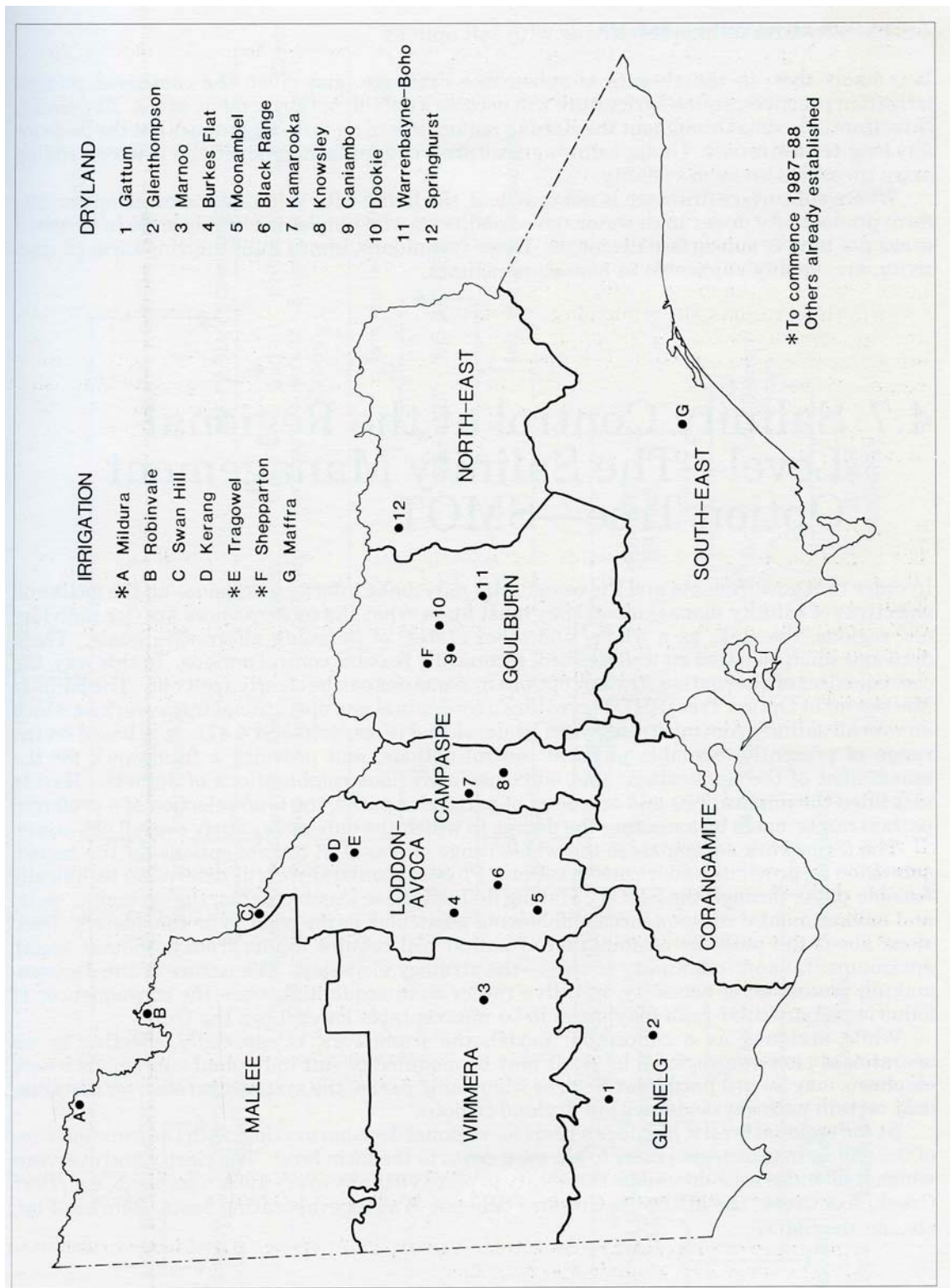


Figure 4.45 – Victorian salinity regions

4.7 Salinity Control at the Regional - Level the Salinity Management Option Tree SMOT

In order that governments and the community may make informed decisions on the goals and objectives of salinity management, they must know what the consequences are (for individuals, regions, the state as a whole, and other states) of pursuing alternative goals. These decisions must be based on well-defined, technically feasible control options. In this way, the consequences of alternative strategy options or scenarios can be clearly spelt out. The Salinity Management Option Tree (SMOT) provides a conceptual and operational framework on which an overall salinity control strategy can be developed (Figures 4.46 and 4.47). It is based on the range of presently available physical control options, and provides a framework for the assessment of the implications and outcomes of various combinations of options. Having identified the implications and outcomes of particular paths, the final selection of a preferred path(s) can be made by assessing the degree to which the outcomes satisfy overall objectives.

The framework encompasses the whole range of potential control options for the region, subregion or province under consideration. Physical constraints will determine technically feasible paths through the SMOT. Having defined these feasible paths, the economic, social and environmental consequences of following particular paths need to be considered. Decisions about the path (or combination of paths) will require inputs from economic, social, environmental and community sources - the strategy elements. The nature of the decision-making process is, of necessity, iterative rather than sequential, since the consequences of following a particular path may prove to be unacceptable lower down the tree.

While designed as a conceptual model, the framework is especially effective as an operational plan at a regional level. It may be modified to suit individual subregions where emphasis may lie in a particular field, or where only part of the system operates; for instance, only certain pathways exist within dryland regions.

At the regional level it provides a basis for regional decision making, with the consequences of the effects in that region clear to all, even down to the farm level. The clarity and overview which it provides is evidenced by the ability to place virtually every known scenario, e.g., Barr Creek, Axe Creek, the MRB, the Girgarre Scheme, Wakool evaporating basin, Bourkes Flat, etc., on the SMOT.

4.7.1 Comments on the SMOT

1. The initial division is into irrigation and dryland situations.
2. The major pathways are largely predetermined by physical parameters, independent of social and economic considerations; for instance, the division into local and regional flow systems in the dryland areas.
3. While there is clearly choice in the case of the three basic irrigation pathways (land retirement, subsurface drainage, and on-farm management), the choice in individual cases is limited by the constraints of what is physically possible. For instance, in irrigation districts it is likely that suitable aquifers which will enable a pumping option may underlie only 35% of the irrigated area. Where subsurface drainage is neither available nor economic, it seems inevitable that the eventual options will be between saline agriculture and land retirement.
4. In the case of irrigation regions, disposal to the Murray River is a dominant path-determining step; the implications of a non-Murray disposal policy are clearly spelled out in terms of the remaining two options (Figure 4.48).
5. In the dryland category, only 10-30% of the area has a local flow system amenable to short-term salinity control.

In conclusion, the Salinity Management Option Tree approach enables integration of strategy

elements with technically feasible salinity control options to produce a range of strategy scenarios.

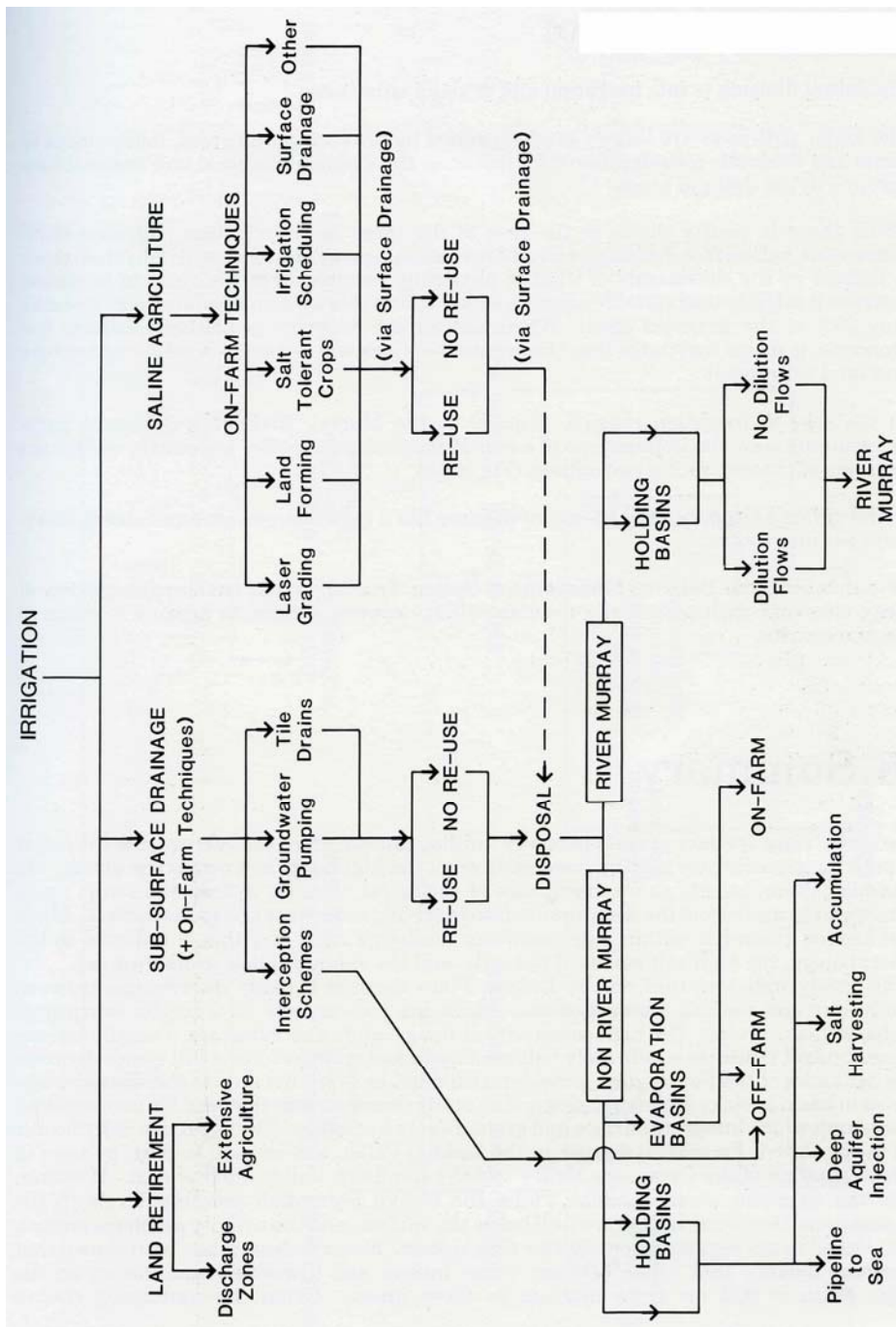


Figure 4.46 – SMOT for irrigation regions

4.8 Summary

"Deep leads" (late Tertiary gravels partially infilling ancient river valleys) provide the major flow path for groundwater moving downbasin from the highlands and across the plains. On the Loddon Plain, largely as a consequence of geological controls, hydrostatic forces cause groundwater to move from the deep aquifer into overlying sediments and to the surface. Much of the Loddon Plain lies within a groundwater discharge zone, and this is reflected in the geomorphology, the high salt status of the soils, and the saline shallow groundwaters.

This study indicates that on the Loddon Plain there is a ready interchange between groundwater and surface water systems, which are two aspects of a single, integrated, downbasin flow system. The high sensitivity of the groundwater system to a small increase in water budget indicates a delicately balanced hydrologic regime. For a full comprehension of the dynamics of the flow regime, consideration must be given not only to the water budget but also to basin geology and morphology. The study demonstrates the need for management and research which integrate surface and groundwater hydrology. The rapid rise in hydraulic head in the Calivil Formation aquifer in the Loddon Valley was echoed, in part, by rises in heads in aquifers of the Campaspe Valley and the Goulburn Valley, further east. However, unlike the situation on the Loddon Plain, the Calivil Formation pressure levels in the Campaspe and Goulburn valleys are well below the surface, and the salinity problems are not, as yet, linked to the regional groundwater flow system. Nevertheless, it has been shown that the critical balance that exists between water budget and hydrologic equilibrium on the Loddon Plain is still an acute problem in those areas. Given the continuing rise in potentiometric head within the regional groundwater system, it can be predicted that a regional groundwater discharge zone will eventually emerge within the central Goulburn Plain and the Campaspe Plain, thus leading to a severe exacerbation of the present salinity problems.

Within the Mallee area of northwestern Victoria, groundwater outcrop occurs in the lower depressions of the landscape. Because the elevation of the regional water table developed in the unconfined Parilla Sand aquifer, can be readily obtained by either piezometers or from levels of nearby salinas, it can be predicted which areas are under immediate threat. In most instances, the recently salinized depressions can be shown to have undergone previous cycles of salinization during wetter periods in the past.

Similarly, regional groundwater pressures are rising in bedrock aquifers which extend in an arc across central Victoria. Here, the aquifers are fractured rock systems, with large salt storages occurring in both the saturated and unsaturated zones. The salinity problem is twofold - land and stream salinization. Similar processes have occurred cyclically in the past in response to climatic changes; it is considered that these saline cycles have, in turn, induced cycles of landscape modification across the highlands: the process here being termed the saline cycle of erosion.

Across northern Victoria, therefore, there is now taking place a fundamental readjustment in the hydrologic equilibrium within the groundwater systems, brought about by increased recharge following the impact of European occupation on what was a previously untouched landscape. While the ultimate extent of these adjustments is not yet known, the direction is clearly shown by geomorphic markers, which represent the effects of previous naturally occurring cycles of higher water budget. Given an understanding of the structural and stratigraphic framework of the basin and the dynamics operating within the flow systems, it is possible to gain an insight into the environmental effects that may have accompanied the hydrologic changes which have occurred during Quaternary times. The traces of these earlier saline cycles, when taken together with the present-day hydrologic trends, may, in turn, provide a basis for the prediction of the eventual extent of the present saline cycle and its future severity.

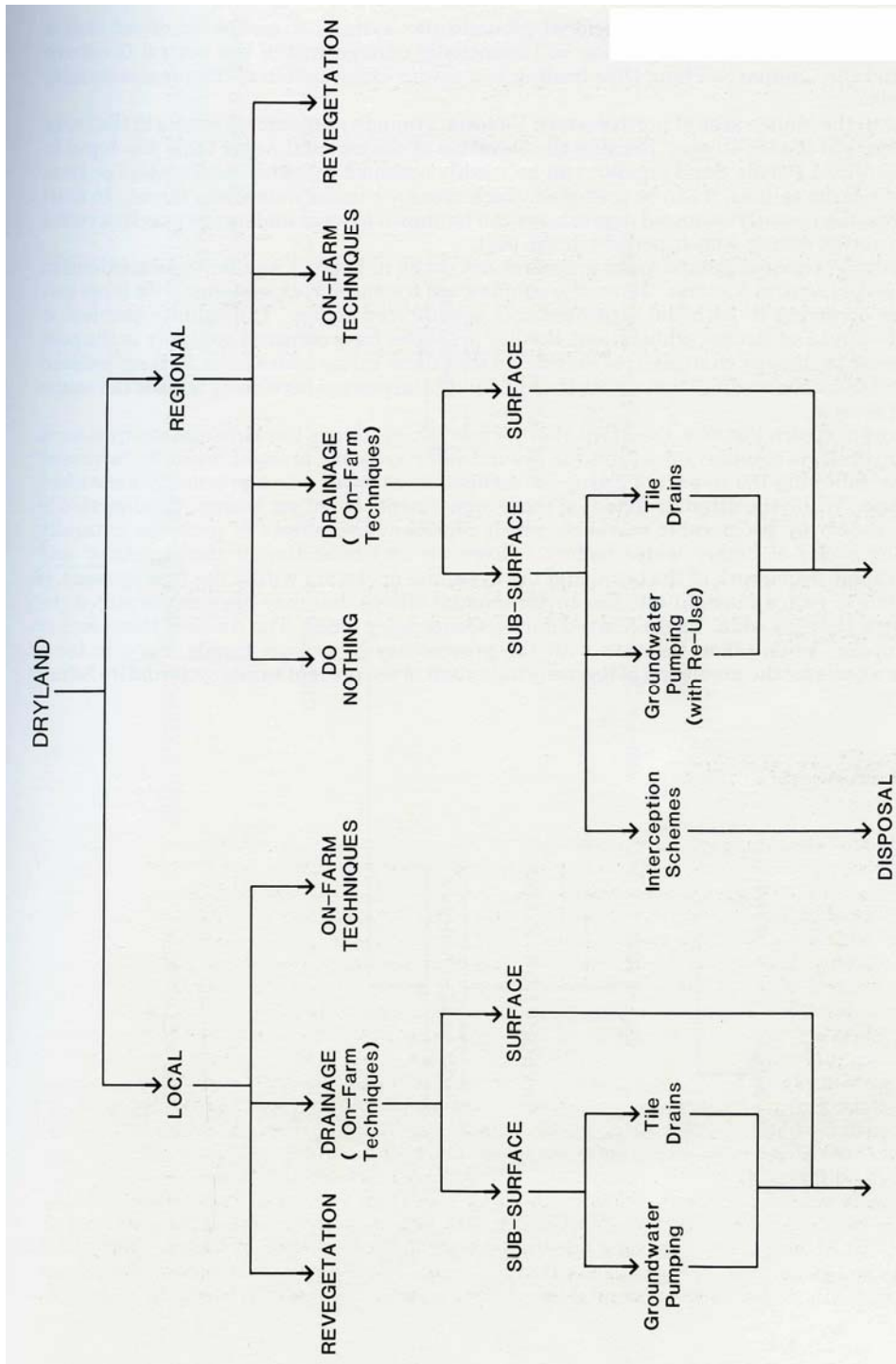


Figure 4.47 – SMOT for dryland regions

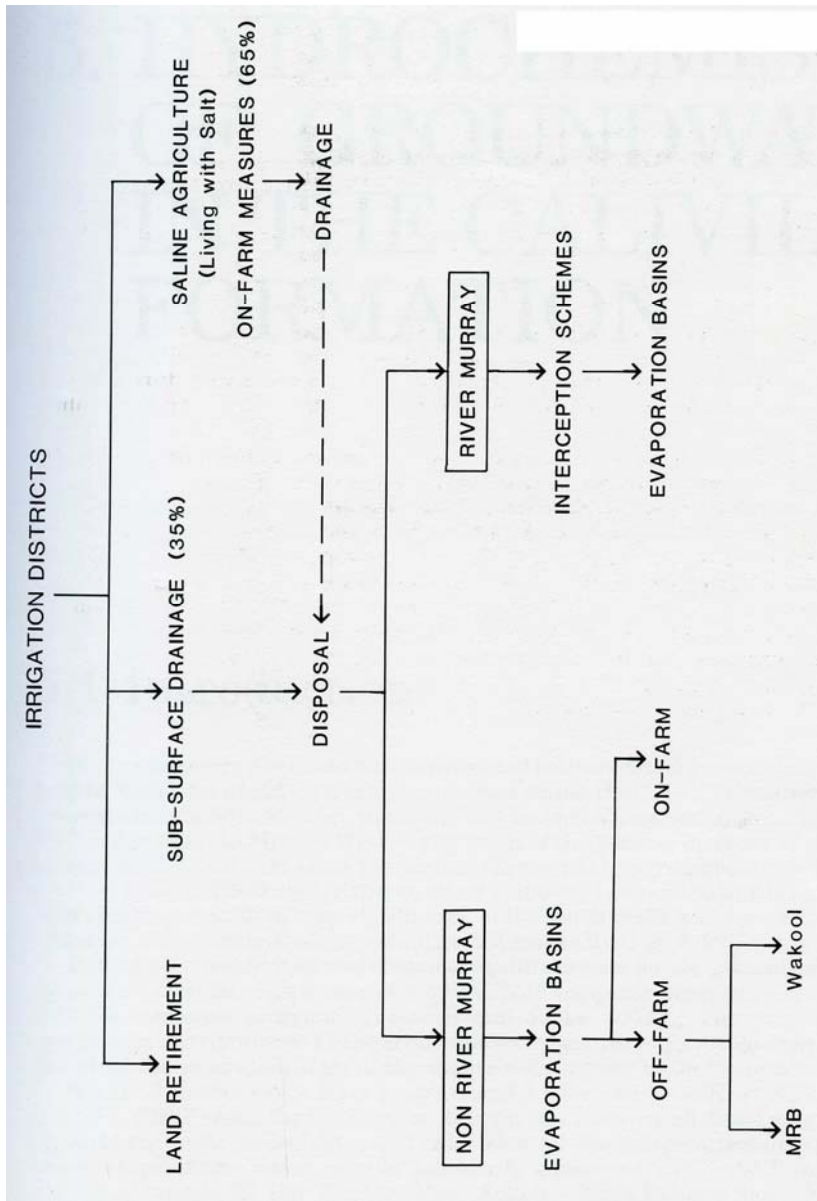


Figure 4.48 – Abbreviated SMOT for irrigation districts

5. HYDROCHEMISTRY OF GROUNDWATER IN THE CALIVIL FORMATION

5.1 Introduction

The hydrochemistry discussed here is restricted to the regional groundwater passing down the Loddon Valley deep lead system and contained within the Calivil Formation aquifer. The area covered extends from Moolort, in the central Loddon Valley within the highlands, out across the Loddon Plain to Mincha West in the north of the plain, a distance of about 150 km. The salinity profile (Figure 5.1) along the section shows that the groundwater within the highlands is about 1,000 to 1,200 mg/l. rising to about 2,300 mg/l across the southern Loddon Plain, but then climbing rapidly to over 10,000 mg/l in the north of the plain prior to the junction with a Murray Valley system where salinities fall markedly (Figure 4.6c).

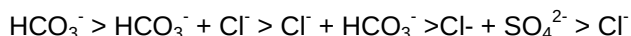
Individual ion activities and chemical equilibria data for the groundwaters of the Loddon Basin were calculated by means of the WATEQF computer program, a revised version of the WATEQ computer program (Truesdell and Jones, 1974). The program calculates the equilibrium distribution of 114 aqueous species in natural waters from their chemical analyses and gives the saturation state of the waters with respect to the various mineral phases.

WATEQF consists of a main program and 5 subroutines - PREP, SET, MODEL, PRINT and SAT. PREP reads the input water data (in this instance all data being provided as mg/l), converts the units to molality, and calculates all the temperature dependent data at the temperature of the water sample arbitrarily chosen as 20°C. SET initializes values of individual species for the iterative Mass Action - Mass Balance loop. MODEL calculates activity coefficients, PRINT prints the results calculated from the aqueous model, and SAT calculates and prints the thermodynamic saturation state of the water with respect to the various minerals considered by the program.

To assist in the examination of the chemical evolution of the groundwaters in the regional Calivil Formation aquifer, the program was run on 38 representative waters from localities along the Loddon Valley, from Glendaruel, south of Moolort in the highlands, to Mincha West in the northern Loddon Plain (Table 5.2).

5.2 Groundwater Classification

The general groundwater sequence along the Loddon Valley follows the classical sequence of Chebotarev (1955), who observed that all groundwaters tend to the composition of sea water. Chebotarev's sequence is



In order to have a quantitative framework to separate the categories, a modified version of Back's (1961) classification of hydrochemical facies has been adopted (Table 5.1).

TABLE 5.1 - Classification of hydrochemical facies*

Hydrofacies	Percentage of constituents in equivalents per million			
	Ca+Mg	Na+K	HCO ₃ +CO ₃	Cl+ SO
Cation facies:				
Calcium-magnesium	90 - 100	0 - 10		
Calcium-sodium	50 - 90	10 - 50		
Sodium-magnesium	10 - 50	50 - 90		
Sodium-potassium	0 - 10	90 - 100		
Anion Facies:				
Bicarbonate			90 - 100	0 - 10
Bicarbonate-chloride-sulphate			50 - 90	10 - 50
Chloride-sulphate-bicarbonate			10 - 50	50 - 90
Chloride-sulphate			0 - 10	90 - 100

Modified from Back (1961)

Chemical Analyses

The chemical analyses of water samples used in this study were carried out by the Department of Minerals and Energy's Chemical Branch at the Victorian State Laboratories, under the direction of Mr R Mealan, Deputy Chief Chemist. The following is a summary of the techniques used for chemical analysis.

TOTAL SOLUBLE SALTS: BY SUMMATION OF CATIONS AND ANIONS

Chloride:	Potentiometric titration with silver nitrate
Carbonate& bicarbonate:	Potentiometric titration with standard acid, using a pH electrode
Sulphate:	1. Precipitation as barium sulphate, determined gravimetrically 2. Turbidimetric, with barium chloride
Nitrate:	Ultra violet absorption
Calcium:	Atomic absorption
Magnesium:	Atomic absorption
Sodium:	Atomic absorption
Potassium:	Atomic absorption
Soluble iron:	Atomic absorption
Total iron:	Atomic absorption
Silica:	Colorimetric, using ammonium molybdate
Bromide:	Volumetric, oxidation to bromate, then titration with sodium thiosulphate

SG: Comparing mass of a known volume with distilled water

pH/Eh: While pH values were automatically recorded in the laboratory, where consideration is given to pH/Eh relationships - as throughout Chapter 7- measurements were taken in the field using a Titron k-240 portable pH/Eh meter. This meter contains a combined pH electrode, which reads on two scales, 0-8 and 6-14. Two electrodes were used for Eh measurements - platinum electrode and calomel reference electrode.

TABLE 5.2 - Hydrochemistry of Calivil Formation waters (mg/l)-Loddon Valley

Bore No.	TDS	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	Na ⁺	K ⁺	Ca ²⁺	Mg ²⁺	SiO ₂	pH
Glendaruel 34	605	80		284	80	4	27	26	104	8.15
Neereman 10001	957	322	48	281	210	6	29	46	15	7.3
Neereman 10016	1049	273	1896	21	302	7	22	9	19	7.2
Moolort	902	216	90	316	136	6	39	61	38	7.85
Moorlort 8008	1026	260	96	342	231	-	27	43	23	7.4
Moolort 8009	1050	283	97	323	249	-	23	49	24	8.3
Baringhup 8002	866	259	27	329	153	-	45	53		7.5
Baringhup 10003	1232	441	79	305	256	8	42	66	36	8.0
Baringhup 10006	1315	456	68	357	238	8	58	85	45	7.5
Laanecoorie 51	1368	502	95	336	304	-	47	74	10	7.1
Laanecoorie 10001	1162	430	35	287	270	7	38	53	32	8.1
Bridgewater 17	1017	414	23	305	281	-	14	45	3	8.6
Bridgewater 8007	1161	480	76	305	331	-	27	53	26	7.0
Bridgewater 6	1196	460	67	282	306	-	27	51	1	7.5
Bridgewater 8008	1476	646	1038	20	390	-	27	66	35	6.9
Yarraberb 8005	1247	444	66	313	277	8	40	56	43	7.75
Yarraberb 10002	1382	546	79	282	357	9	38	49	22	8.5
Yarrayne 10010	1440	565	88	289	345	7	39	62	45	7.6
Pompapiel 1	1373	590	24	302	356	8	21	62	10	8.2
Pompapiel 8001	1688	729	110	526	418	8	43	75	32	7.8
Pompapiel 8013	1928	842	165	226	526	8	45	74	42	8.3
Janiember East 8004	1435	582	88	269	352	7	35	61	21	7.7
Janiember East 8014	1496	612	96	268	372	7	35	63	43	7.9
Janiember East 8008	1560	673	102	238	387	7	39	69	45	7.95
Janiember East 8007	1595	717	94	226	384	5	49	75	45	7.4
Janiember East 16	1572	769	39	218	440	7	17	78	3	8.3

Bore No.	TDS	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	Na ⁺	K ⁺	Ca ²⁺	Mg ²⁺	SiO ₂	pH
Talambe 8001	2096	977	153	245	536	10	55	96	24	8.0
Calivil 8004	2155	999	162	256	552	9	58	97	22	7.6
Calivil 8003	2197	1034	146	248	570	9	63	105	22	7.2
Calivil 5	2266	1057	167	260	587	9	63	105	18	7.5
Calivil 9	2198	1039	165	244	570	9	63	103	5	8.2
Calivil 3	2303	1079	181	253	588	10	64	106	21	7.4
Calivil 8002	2307	1090	133	296	602	10	59	105	12	7.9
Mologa 1	6248	3350	357	322	1815	17	127	253	8	7.8
Mologa 2	6312	3330	407	317	1827	19	140	264	8	7.7
Mologa 8001	6050	3135	357	330	1804		107	213		7.3
Mincha West 12	9119	5112	678	70	2560	14	262	417	5	7.3

5.3 The Chemical Evolution of Groundwater in the Calivil Formation

Upstream of Bridgewater, recharge waters percolate downward through valley flow basalts (Chapter 4.1) prior to reaching the pebbly wash and quartz drift sand of the Calivil Formation. Of the basalts, the most common rock types are alkali olivine basalt and olivine tholeiite, for which chemical analyses are given in Table 5.3.

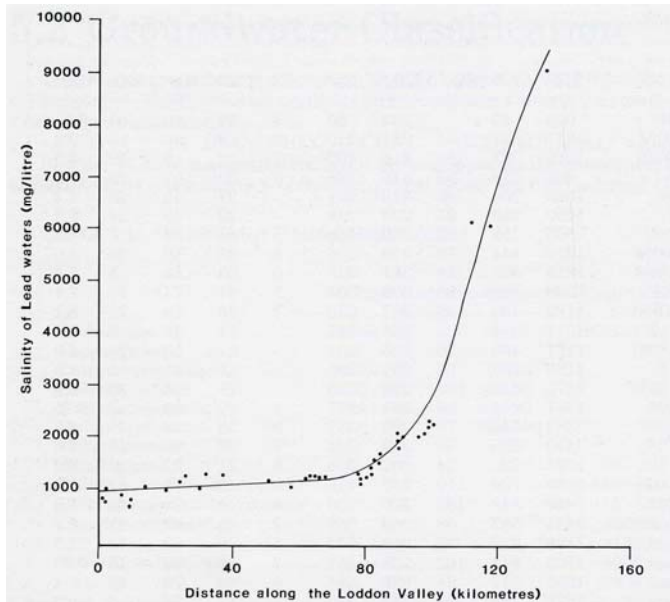


Figure 5.1 - Salinity variations in the Calivil Formation of the Loddon Valley between Moolort and Mincha West. The rapid rise in salinity commences virtually at the hinge line between the regional groundwater recharge and discharge zones in the Pompiapiel district (see also Figure 4.6c).

TABLE 5.3 - Analyses of basalt types found in the upper Loddon catchments*

	Analysis 1	Analysis 2		Analysis 1	Analysis 2
SiO ₂	42.30	45.31	Orthoclase	8.51	8.27
TiO ₂	3.21	2.95	Albite	18.68	25.30
Fe ₂ O ₃	4.39	3.74	Anorthite	21.80	26.13
FeO	7.39	7.54	Diopside	17.12	12.10
MnO	0.18	0.17			
MgO	10.77	7.66			
CaO	9.46	8.87			
Na ₂ O	2.81	2.99			
K ₂ O	1.44	1.40			
P ₂ O	0.55	0.43			
H ₂ O	0.74	0.46			

Analysis 1 - Average of 28 alkali olivine basalts

Analysis 2 - Average of 37 olivine tholeiites

*(Data from Ellis and Ferguson, 1976)

In the upper reaches of the Loddon Valley, the newly recharged groundwaters in the Calivil Formation are of the calcium-sodium bicarbonate-chloride-sulphate hydrofacies. The waters reflecting the influence of the basalts are either saturated or near to saturation in magnesite, calcite, and dolomite (Figure 5.3); therefore, initially, on passing downbasin, the increased salinity is accompanied by relative rises in Na⁺, Cl⁻ and SO₄²⁻, while Ca²⁺, Mg²⁺ and HCO₃⁻ remain fairly constant. Using percentage as a basis, plots of anions and cations show an initial rapid falloff in

the latter ions, while the former show a commensurate rise (Figure 5.2). Thus, even a slight rise in salinity results in the development of a sodium-magnesium chloride-sulphate-bicarbonate hydrofacies while still well within the highland tract where chlorinity values are only slightly above 200 mg/l. This stage has already been reached at Moolort, and, on passing further downbasin, the trend continues until a chloride-sulphate hydrofacies develops in the northern Calivil district on the Loddon Plain. In the case of the cations, the $\text{Ca}^{2+} + \text{Mg}^{2+}$ percentage remains less than 50%, being about 20% to 30% over most of the section. Under Back's (1961) classification, this is a $\text{Na}^+ - \text{Ca}^{2+}$ water; however, given the ascendancy of the Mg^{2+} ion over the Ca^{2+} , it can be better considered as a $\text{Na}^+ - \text{Mg}^{2+}$ water, and this category is preferred in Table 5.1.

While, in the uppermost regions of the Loddon Valley, the cation ratios undergo rapid changes as the Na^+ percentage rises relative to Ca^{2+} and Mg^{2+} , this lasts for only a short distance on passing downbasin before leveling off. Indeed, perhaps the most striking feature of the groundwater chemistry is seen in the cation percentages (Figure 5.2), which show that by about chlorinities of 600 mg/l (about 1,400 mg/l, TDS), that is, in the vicinity of the Pompapiel district, the groundwaters have virtually reached an equilibrium situation, having constant cation ratios, which change only slightly on crossing the Loddon Plain. The concomitant rise in Ca^{2+} and Mg^{2+} to that of Na^+ , as salinity increases, occurs despite the saturation or near saturation of the groundwaters with respect to the carbonate phases. At the same time, there is a steady decline in the percentage HCO_3^- and CO_3^{2-} species (shown only as HCO_3^- in Table 5.2), suggesting the ongoing precipitation of the carbonate phase, or its nonsolution.

The uniform groundwater character from Pompapiel northward shows that cation ex-change has virtually no effect within the Calivil Formation beneath the Loddon Plain, partially reflecting the lithology and mineralogy of the Calivil Formation, which consists, in this region, of clean, coarse-grained quartz gravel and sand, with no lithic fragments other than quartz. (Only one nonquartz fragment was obtained from the Calivil Formation during the drilling program undertaken for this investigation.)

In contrast to the cation ratios, which show little change after chlorinities of 600 mg/l, the anions show a marked increase in chloride percentage on passing down the basin and a concomitant decrease in bicarbonate (Figure 5.2). Although initially somewhat irregular, there is a general trend for the sulphate to remain constant, within a range of 3-14% overall, and from 7-11% beyond Bridgewater. The rate of decline of the bicarbonate + carbonate percentage is rapid until chlorinities of about 1,300 mg/l are reached (approximately 2,700 mg/l, TDS), after which, although still continuing to decline, the rate of decline levels off.

Given the above trends, it can be shown that the chloride-cation ratios, although undergoing initial rapid rises (less in the case of chloride-sodium), become more constant by chlorinities of about 800 - 1,000 mg/l. This stabilization is reached in the Pompapiel to Calivil region on the Loddon Plain. Apart from rises in the Cl^- percentage and falls in the HCO_3^- percentage, the character of the groundwater is relatively uniform beyond Calivil, with only salinity showing any significant changes as it rapidly rises from 2,300 mg/l to over 6,000 mg/l in the parish of Mologa immediately to the north.

Perhaps the most significant aspect of the trends, in terms of the downbasin evolution of the groundwaters, is that by the time salinities have reached values of only about 2,000 mg/l the waters are already plotting close to the field of seawater (Figure 5.4a). Therefore, by this early stage, the groundwaters have already obtained a distinct oceanic character, which is only slightly modified on further concentration. The trend towards an oceanic character is also present in groundwaters contained in shoestring sand aquifers in the overlying Shepparton Formation of the southern Loddon Valley, and in waters of a Murray Valley lead system in the northernmost regions of the Loddon Plain (Figures 5.4b and 5.4c). While it could be argued that a connate seawater influence may have been present since the early Pliocene sea transgressed this far inland, it is not the case with the waters from the upper Loddon Valley, which come from a continental setting well beyond the limits of the Pliocene sea. Similar trends occur in the

groundwaters from the Campaspe Valley (Figure 5.5a).

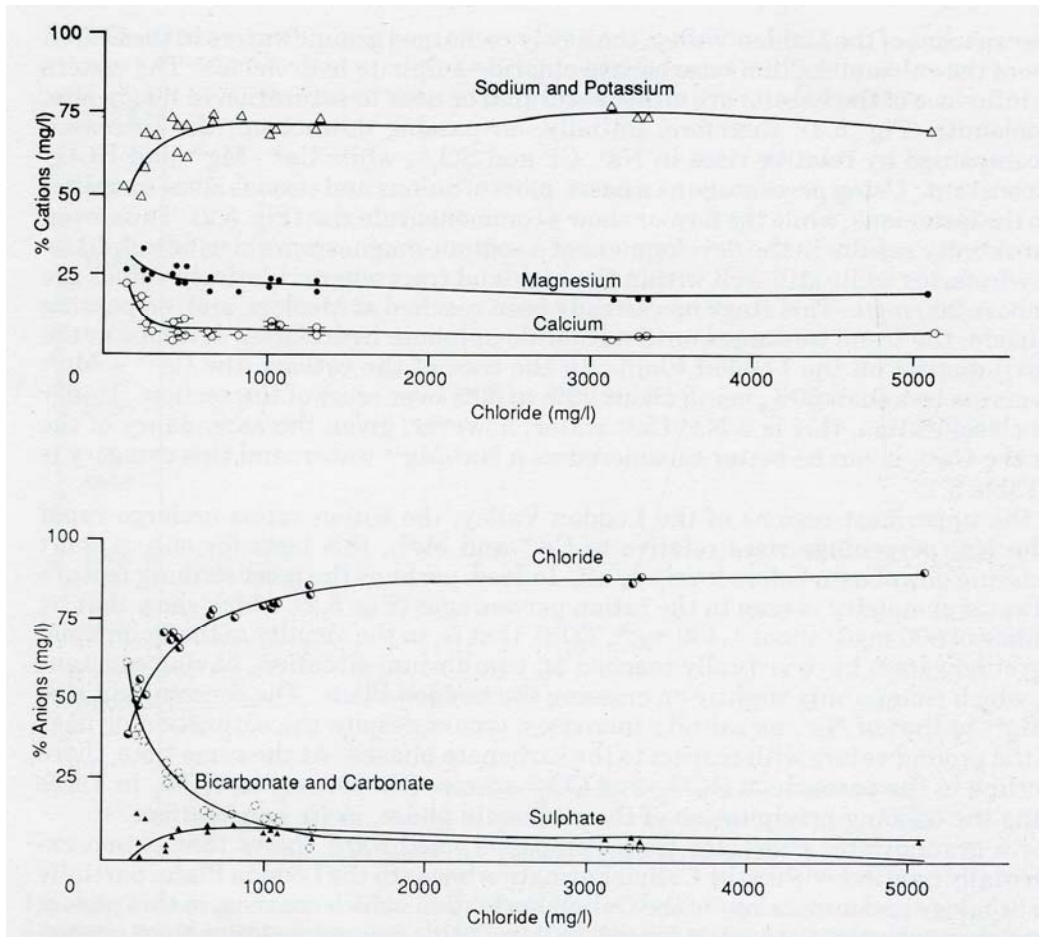


Figure 5.2 - Cation and anion percentages versus chloride concentration - Calivil Formation, Loddon Valley (between Moolort and Mincha West)

The early attainment of an oceanic character by the regional groundwaters while still well within a continental setting is an important consideration in any discussion of the origin of the saline oceanic-type groundwater in the marine Parilla Sand aquifer (Figure 5.5b), which is the downbasin continuation of the Calivil Formation aquifer in terms of regional groundwater flow (Chapter 4.2).

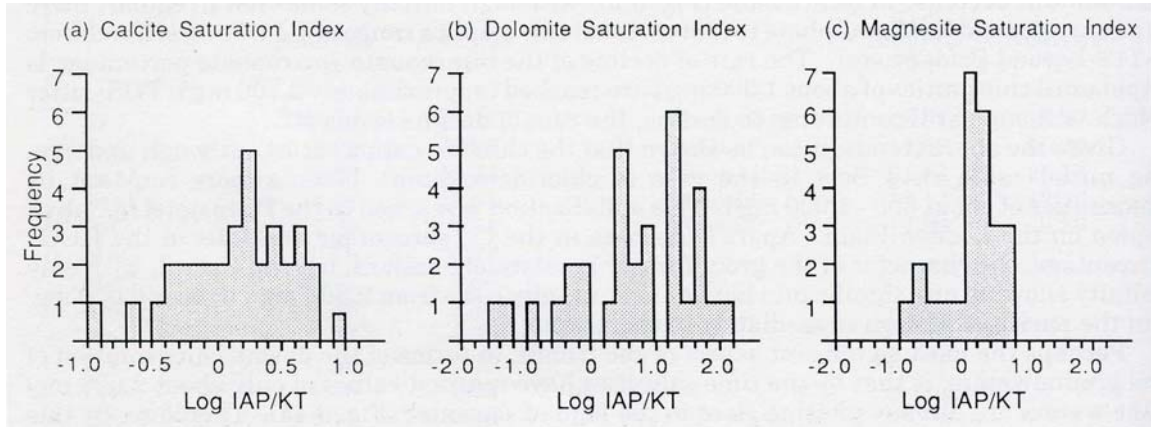


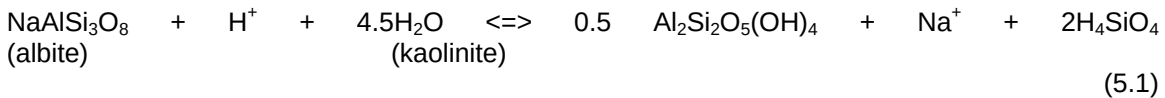
Figure 5.3 – Carbonate phase saturation indices – Calivil Formation, Loddon Valley

5.4 Silicate Equilibria in the Calivil Formation Aquifer

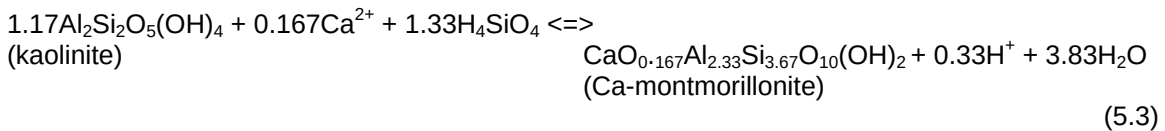
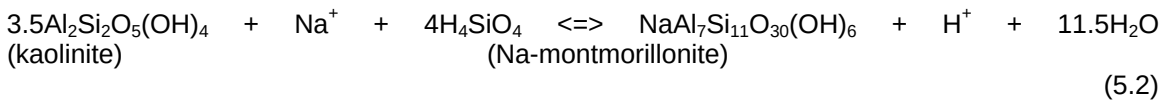
5.4.1 Evolutionary Trends in Silicate Equilibria

The early attainment of cation equilibrium along the flow path causes groundwaters which occur in the highland tract to be already essentially chemically mature: they undergo little further evolution on passing downbasin. A similar evolutionary trend towards chemical maturity in silicate equilibria was suggested by Helgeson et al., (1969). This process commences with the incongruent dissolution of aluminosilicate minerals by CO₂ enriched groundwaters, which convert feldspars and micas to kaolin, with a corresponding increase in Na⁺, Ca²⁺, K⁺, Mg²⁺, and H₄SiO₄ relative to H⁺ (Stumm and Morgan, 1970).

For example:



Following further weathering and the increased availability of silica and cations, the kaolinite can then be converted into montmorillonite as follows



The implication is that the sediment-water system is closed, and, therefore, there is no loss of ions from the system. However, in natural systems, when feldspar-rich igneous rocks are infiltrated by aggressive recharge waters rich in CO₂, they are leached of cations and silica. While the initial response is to form gibbsite, the amount of silica cations released is normally more than sufficient to convert the gibbsite to kaolinite (Stumm and Morgan, 1970). As the process proceeds, the cation and silica concentrations steadily increase to a point where the kaolinite starts converting into montmorillonite. Predominance diagrams illustrating the typical progress of mineral stability in the weathering process are shown in Figure 5.6 (from Stumm and Morgan, 1981).

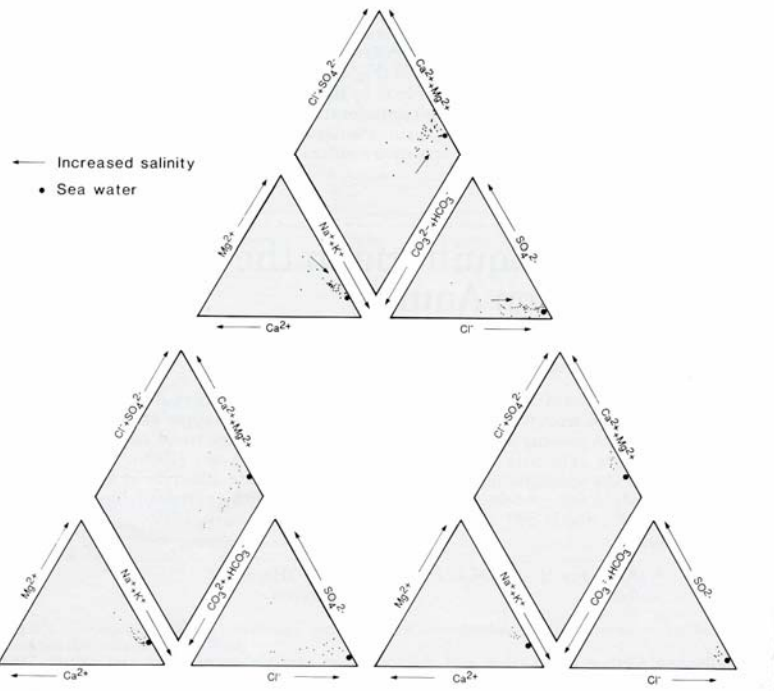


Figure 5.4a - Hydro-chemistry of the Calivil Formation - Loddon Basin

Figure 5.4b - Hydro-chemistry of the Shepparton Formation on the southern Loddon Plain

Figure 5.4c - Hydro-chemistry of Calivil Formation - Murray Valley lead on the northern Loddon Plain

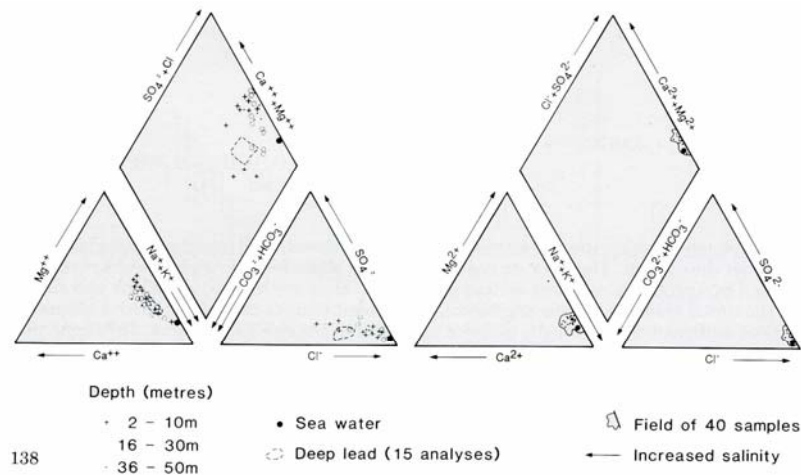


Figure 5.5a - Hydro-chemistry of Campaspe Valley groundwaters (from Tickell, Unpub.)

Figure 5.5b - Hydro-chemistry of Parilla Sand

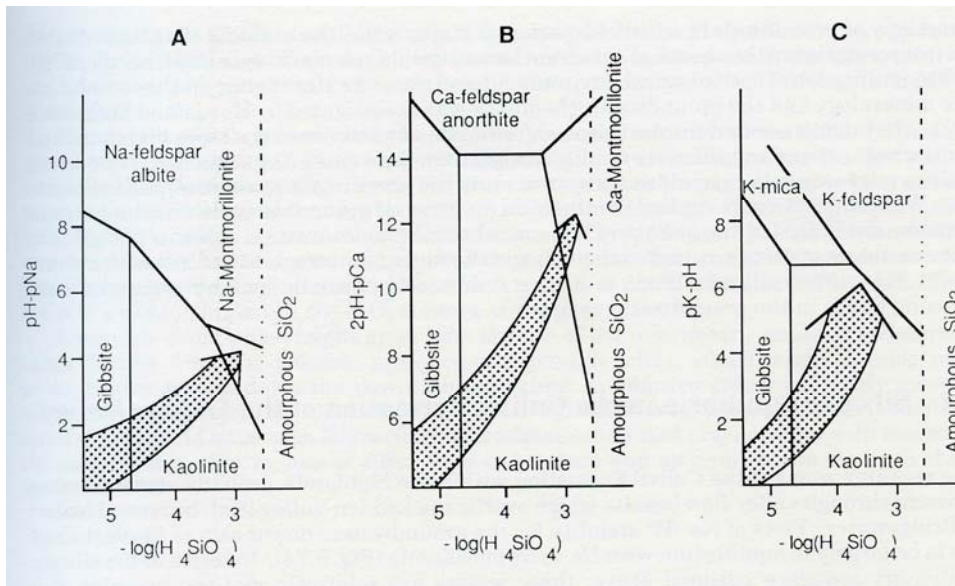


Figure 5.6 - Predominance diagrams illustrating the stability relationships between gibbsite, kaolinite, montmorillonite and feldspars. Arrows indicate hypothetical paths taken during the weathering process (after Stumm and Morgan, 1981).

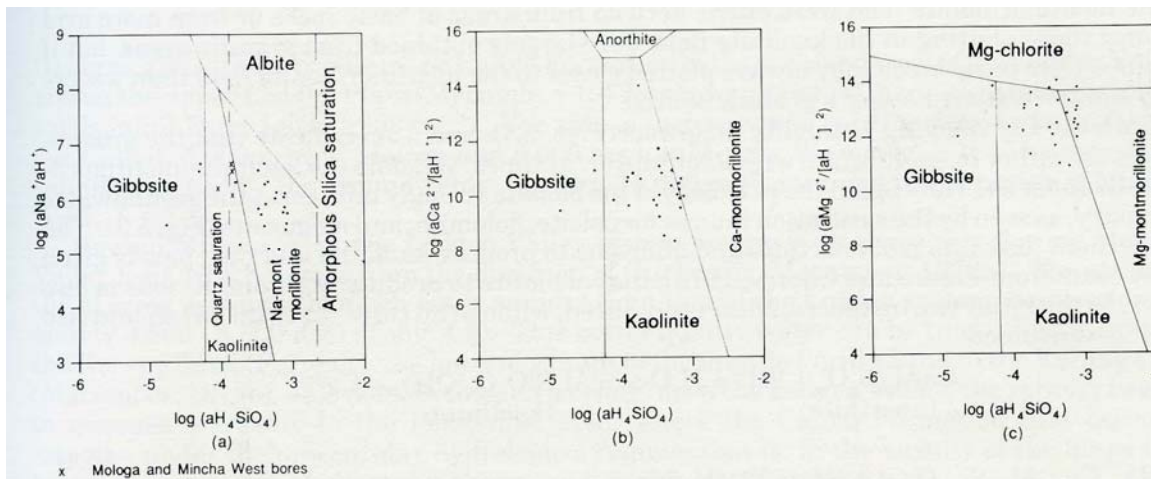


Figure 5.7 - Silicate stability diagram, Calivil Formation, Loddon Valley (at 25° and 1 bar)

Groundwaters in equilibrium with the solid phase minerals will show similar characteristics, which maybe demonstrated by means of activity diagrams (Garrells and Christ, 1965). The boundaries between the differing mineral phases are based on thermodynamic data; however, the constants used may vary between individuals, which in turn leads to differences in the position of the boundaries. The stability boundaries chosen for use in this study are taken from the work of Nesbitt (1977). Slightly different boundaries are shown in Figure 5.6 from Stumm and Morgan (1981), while Tardy (1971) has a further set of boundaries. The stability boundaries are based on pure minerals and, therefore, do not accurately represent the real systems in nature. Nevertheless, the equilibrium data provides a useful insight into the interactions of groundwaters with feldspars and clays, while the stability diagrams enable some interpretation of the chemical data from hydrological systems (Freeze and Cherry, 1979).

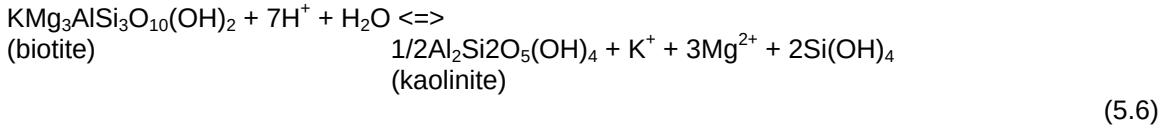
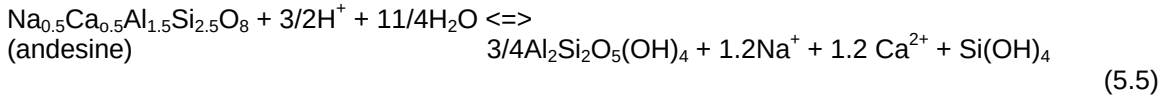
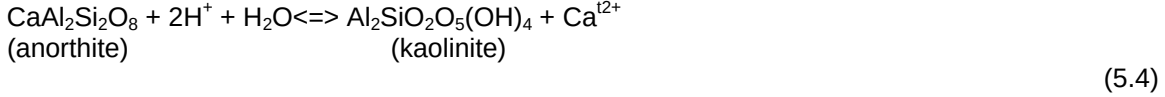
The significance of both physical environment and groundwater regime on the coevolution of the mineralogy and the groundwater chemistry was investigated by Kovda and Samoilova (1969). They demonstrated the preferential formation of kaolinite in the more humid upland regions where cations and silica are readily flushed from the weathering sequence. The cations and silica migrate via the groundwater systems into the lower regions, where montmorillonite forms. Wallick (1981), who studied the chemical evolution of groundwaters in a drainage basin in Canada, noted that in the more arid regions, where the concentration of Na^+ is high (due to the precipitation of calcite, cation exchange, and the higher ionic mobility of Na^+ with respect to Ca^{2+}), Na-montmorillonite tends to be the stable solid phase in lowland areas and with increasing depth in the groundwater system.

5.4.2 Silicate Equilibria in the Calivil Formation of the Loddon Valley

In the recharge zones of the Calivil Formation within the highlands, groundwater infiltrates downward through valley flow basalts which overlie the Loddon Valley lead, between Moolort and Bridgewater. Plots of Na^+/H^+ stability for the groundwaters downbasin of Moolort show them to be largely in equilibrium with Na-montmorillonite (Figure 5.7a). In terms of the silicate evolutionary sequence outlined above, these waters are relatively mature; however, this apparent maturity probably arises directly from influence of the basaltic recharge areas. The groundwaters are oversaturated with quartz, which was probably derived from the aluminosilicates in the recharge areas. It was shown by Tardy (1971), who sampled 450 waters from a number of countries having differing geological and climatic regimes, that waters which plot in the montmorillonite field were either derived from areas of basic rocks or from more arid regions; those plotting in the kaolinite field were largely obtained from granitic areas, but if obtained from basic areas they always plotted closer to the montmorillonite field than waters from similar regions having a

granitic source.

Both the Ca^{2+} and Mg^{2+} stability diagrams (Figures 5.7b and 5.7c) indicate that the groundwaters are either in equilibrium with kaolinite or, at times, straddle the kaolinite-montmorillonite boundaries. Here again the petrology of the basalts strongly influences the groundwater chemistry, as seen by the saturation indices for calcite, dolomite, and magnesite (Figure 5.3). The incongruent dissolution of anorthite and andesine to produce kaolinite and Ca^{2+} ions is given below (data from Freeze and Cherry, 1979); that of biotite to produce Mg^{2+} and K^+ ions is also given. In the latter two instances, silica is produced, while in all three cases the pH should rise as H^+ is consumed.



While the silicate stability diagram indicates that the groundwaters are in equilibrium with Na-montmorillonite in the recharge areas of the highlands and the upper parts of the plains, in the lower regions of the flow path, where the Calivil Formation trench is incised into deeply weathered (kaolinized) Palaeozoic sediments, the chemical equilibrium swings toward the kaolinite field (Figure 5.6a). An indication of this trend can be seen in the silica content of the groundwaters, which falls away sharply in the northern parishes of Mologa and Mincha West (Table 5.2).

It can be seen therefore, that despite the general evolution of the Calivil Formation groundwaters toward uniformity and maturity, even at an early stage along the flow path the trends within the silicate stability do not show any comparable pattern but appear to relate more to the mineralogy of the area through which they are passing. While the presence of high SiO_2 and Na^+ content is essential to the development of groundwaters in equilibrium with Na-montmorillonite in the upper and central regions of the flow path, the movement of the equilibrium toward kaolinite in the lowermost regions is not due to a lack of Na^+ but instead reflects a diminishment in the SiO_2 content of the groundwaters.

A possible error which might arise from the use of the colorimetric analysis technique for silica derives from the possible presence of polymeric silica, which might develop in the groundwater system down the flow path with time; in addition colorimetry only measures silica above the dimer (R.J. McLaughlin, pers. com.). However, should polymerization have occurred, it would have been followed by aggradation and rapid precipitation with essentially the same result, that is, loss of silica from the system and an equilibrium closer to that of kaolinite.

5.5 Salinity Profile down the Loddon Valley

General: The salinity profile in the Calivil Formation in the highland tract of the Loddon Valley and on the upper Loddon Plain (Macumber, 1978c) shows a gradual increase in total dissolved solids, until Pompapiel (1,250 mg/l). The rate of increase then rapidly rises, to be 2,200 mg/l at Calivil, 6,000 mg/l at Mologa and 9,000 mg/l at Mincha West (Figure 5.1). The maximum salinity reached by the groundwaters is about 15,000 mg/l, recorded in the parish of Mincha West.

Beyond Mincha West, the Loddon Valley system merges with a westerly flowing Murray Valley lead, which passes from the direction of Gunbower (Macumber, 1978a). The water in the Murray system is of much lower salinity than that of the Loddon system, being approximately 3,500 - 6,500 mg/l (Table 4.2). This better quality water can be traced westward into the Kerang Sand (Figure 4.6c), the lowermost unit of the marine Parilla Sand in the Kerang area (Macumber, 1978b). It is notable that, on passing down the Loddon Valley, the salinity begins to increase markedly in the Pompapiel area, where the Calivil Formation first becomes artesian under the present-day hydrological regime, that is, in the vicinity of the hinge line between regional groundwater recharge and discharge (Chapter 4.1.11).

5.6 Factors which Influence Groundwater Salinity

Mechanisms which control the chemistry of groundwaters are well known, and those which influence the Murray Basin groundwater systems have been previously listed by Lawrence (1975). Certain of these processes have been referred to in more detail in the discussion above, while others are described elsewhere in this report. A brief resume of these processes is given below:

- i. The addition of atmospheric salts, which fall with rainwater.
- ii. Addition of carbonates, formed by the solution of CO_2 which is produced in the soil zone by the decay of organic matter and the respiration of microorganisms.
- iii. The addition of oceanic connate water. (This does not appear in the fluvial sequences, away from marine sequences.)
- iv. Dissolution of evaporite minerals such as gypsum, calcite and halite.
- v. Oxidation of pyrite to give sulphate, and the reduction of sulphate to sulphide, in the presence of sulphate-reducing bacteria.
- vi. Cation exchange on clays.
- vii. Ultrafiltration, or salt sieving, by semipermeable clay membranes.
- viii. Chemical weathering of silicate minerals (as discussed above).
- ix. Concentration by evaporation or evapotranspiration.
- x. Residence time. It follows that a longer residence time permits the above listed processes to proceed further, thus increasing salinities.

Undoubtedly, many of these processes contribute toward the chemical character of Calivil Formation waters at some stage along their evolutionary tracts and are responsible for the trends described by Chebotarev (1955), whereby groundwaters tend toward a seawater composition while at the same time steadily increasing in salinity. However, in the case of the Calivil Formation waters of the Loddon Valley, the rapid salinity increase occurs well after chemical equilibrium has been reached and after many of these processes (for example, cation exchange) have virtually ceased to be active. Of the processes listed, most may be discounted as contributing to the salinity increase, leaving only two or three for which discussion is warranted, and these are considered below.

5.6.1 Residence Time

It is generally believed that groundwater salinity may simply reflect residence time in the aquifer, that is, sluggish waters and long flow paths lead to higher salinities. The main basis for this opinion is that the above listed processes have continued to be active over a considerable time and for a significant part of the flow path. The effects of residence time are thus reflected by the evolutionary trends of the groundwater chemistry as well as by the gross salinity increases. Yet, this broad generalization does not explain the sudden salinity increase in the aquifer well after chemical equilibrium has already been attained. Indeed, along its entire length the clean channel-fill quartz gravel of the aquifer maintains a moderately high hydraulic conductivity, ranging from 85 m/d at Bridgewater where the salinities are only 1,200 mg/l, to 43 m/d at Mincha West where the salinities are over 9,000 mg/l. Therefore, despite a gradual decrease in hydraulic conductivity, the flow system maintains a fairly high velocity throughout, giving a calculated flow time from Bridgewater to Mincha West of less than 3,000 years (the hydraulic gradient over this interval being about 0.6 m/km). Individual flow velocities at Bridgewater, Calivil, and Mincha West are given in Table 5.4. These values are calculated from the Theis equation, using a porosity value of 0.4, the maximum likely value. For porosities of 0.2, perhaps a better average porosity, the velocities would double and the residence time halve. The hydraulic conductivity values are from

Table 4.1, Chapter 4.2.

TABLE 5.4 - Aquifer parameters and flow velocities - Calivil Formation of the Loddon Plain

Station	Hydraulic Conductivity (m/d)	Hydraulic Gradient ($\times 10^{-3}$)	Velocity (m/year)
Bridgewater	85	0.86	67
Calivil	49	0.55	25
Mincha West	43	0.38	15

5.6.2 Ultrafiltration

The concentration of groundwater by a process of pressure-induced filtration through semipermeable clay membranes (ultrafiltration) has been reviewed by White (1965) and Berry (1969). The process is discussed in some detail in Chapter 12.16.2.

Because the initial increase in the rate of rise of salinities occurs in the part of the aquifer where the groundwater system is either artesian or lies in an artesian transition zone, and the highest rate of salinity increase occurs in the Calivil-Mincha West region, where the overlying aquitard is predominantly clay (and at its most effective), then further consideration must be given to ultrafiltration as a process which might explain the salinity pattern observed in the Calivil Formation aquifer.

Based on observations of downaquifer groundwater salinity increases in three areas (the mid-Continental area, the Woodbine Sandstone of Texas, and the Illinois Basin), Bredehoeft et al. (1963) constructed a model to explain the rapid salinity buildup in aquifers which discharge into closed basins. They put forward a process whereby under artesian conditions, water passing down an aquifer system may discharge upward through an argillaceous aquitard, which acts as a semipermeable membrane permitting the passage of formation waters but retarding the transfer of dissolved ions. The ions are thus retained in the aquifer and cause a salinity build up. The degree of ionic concentration in the aquifer is dependent on the efficiency of the clay membrane overlying the aquifer. Bredehoeft et al. give a series of graphical solutions to the downbasin salinity increases for varying membrane efficiencies, based on the formula

(5.7)

C_w = concentration at distance w from centre of basin
 C_i = concentration of recharging waters
 r = the radius of the basin
 w = distance of point at which concentration is sought
 $(1-e)$ = membrane efficiency

It would seem not possible to apply the formula directly to the Loddon Valley, since the aquifer system although wedging to the north is open-ended at its downstream end. However, as stated by Bredehoeft et al., an almost infinite variety of geological considerations violates the mathematical model. The Loddon Valley as a hydrological basin departs from the model at both its upstream and downstream ends. In the former instance, the system is overlain by, at best, an aquitard, while, in the latter case the basin is not closed but instead only wedges to the north before becoming tributary to the Murray system.

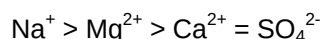
Nevertheless, given these limitations, there is a surprising degree of agreement in salinity values obtained from the Theoretical Concentration Linear Head Model, where,

and those occurring along the Loddon Valley between Bridgewater and Mincha West. Here, the basin radius is taken as being 198 km (passing from Ballarat on the divide due north to the Murray River). The initial concentration (C_i) used for the calculations is 500 mg/l, a figure that approximates the average salinity in the headwater catchment areas. The membrane efficiency is taken as 1. However, where clay membrane sieving occurs it should be accompanied by selective ion filtration.

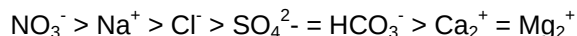
5.6.3 Selective Ion Filtration

A membrane sieving hypothesis was adopted by White (1965) to explain the chemistry of certain North American saline groundwaters. Examples of membrane filtered and membrane concentrated connate waters are given by White, and the rates of migration and the degree of chemical change are considered by him to be dependent on the attendant physical and chemical environments. The initial driving force is seen by White as being lithostatic pressure, after which meteoric waters provide the driving energy under certain circumstances of topography, structure and lithology. Lithostatic pressure is, however, not relevant to the Calivil Formation aquifer, since the situation in the Loddon Valley is not one arising from the compaction of fine-grained sediments containing connate water but, instead, derives from hydraulic heads on meteoric waters. White states that the relative ion mobilities are, in part, consistent with the hypothesis that small single-charge ions are the most mobile and large double-charged ions the least mobile. There is, however, an admission by White that further work is needed in order to fortify or revise the tentative conclusions.

White proposed the order of ion mobility as



Experimental work by Larson (1967) using cellulose acetate membranes suggested that the mobility order was



Unlike the case with the Bredehoeft et al. model, it is difficult to fit the chemical evolution of the groundwater on its passage down the Loddon Valley into the pattern obtained by White. In order to agree with White's hypothesis, decreased Na/Ca, Na/Mg, Na/SO₄, Cl/Ca, Cl/Mg and Cl/SO₄ ratios should occur. But this is not so, since the cation ratios remain roughly constant throughout the system (Figure 5.2). Of the anions, it is the Cl⁻ that shows a tendency to increase preferentially relative to other ions. Like the cations, SO₄²⁻ remains fairly constant, while the HCO₃⁻ shows a marked decrease. An explanation of the latter trend could be contained in White's comment that CO₃²⁻ can migrate across the clay membrane as either uncharged H₂CO₃ or as undissociated NaHCO₃. An order of mobilities, based on the downbasin trends in the Calivil Formation aquifer would be Cl⁻, low; Na⁺, Mg²⁺, SO₄²⁻, Ca²⁺, intermediate; and HCO₃⁻ high.

Therefore, while using only total dissolved solids, a reasonable fit may be had with the mathematical model; however, the hydrochemical data indicates that salt sieving is not a significant process in the salinity increase in the Calivil Formation waters.

5.6.4 Non-Steady State Flow - Salinity Increase in the Discharge Zone

Under the present-day hydrological regime, the Loddon Plain, from Pompapiel in the south through to the Murray River in the north, is underlain by an aquifer which is either artesian or close to being artesian. Between Bridgewater and Mincha West, a distance of about 80 km, there is a loss of over 90% of the water from the Calivil Formation aquifer, partly into the Renmark Group and partly into the overlying Shepparton Formation (Chapter 4.1.11). Over parts of the plain, groundwater discharge to the surface is observed, and elsewhere it is inferred. While some of the outflow occurs as base flow into stream systems and as spring outseepages such as at Bears Lagoon, much of the outflow is deemed to be lost by evapotranspiration and evaporation

from the soil water zone.

While, prior to European settlement artesian conditions would have occurred within the Calivil formation on the Loddon Plain, they were probably not as extensive as they are today. Once irrigation commenced, however, topping up of the excess aquifer storage would have led to a rapid lateral expansion of the regional groundwater discharge zone with the upbasin extension of these conditions. Over much of the Loddon Plain, water tables are now within 1-2 m of the surface, resulting in an epipercolative water regime (cf. Yaalon, 1963), whereby salt accumulates within the soil zone or at the surface (Figure 5.8). About one-third of the area in the mid-Loddon Valley has a high or very high soil salinity. The salinity of the shallow groundwaters is correspondingly high, with salinities ranging from about 10,000 up to 60,000 mg/l. A palaeohydrological regime similar to that existing today was previously documented as occurring at Kow Swamp at the northernmost limits of the Loddon Plain in the early Holocene Period, when high water tables resulted in the loss of trees from the plain (Macumber, 1977).

Indeed, cycles of salinization are seen as recurring events on the Loddon Plain throughout the Quaternary Period whenever high water budgets existed on the plain (Macumber, 1968). The rapid response of the Calivil Formation aquifer in 1973-75 to only several wet years indicates that a very delicately balanced hydrological equilibrium exists in the groundwater system. It also shows that over a period of time, for example, spanning the latter half of the Holocene Period, the aquifer would have responded to any extended wetter or drier events by advances and retreats in the recharge-discharge hinge line. It is plausible to envisage a broad zone in the upper to central Loddon Plain, where the flow system may at one time have a vertical flow component downward, and at another time upward, as was the situation in the Pomppapriel area over the period 1970 to 1980. That is, non-steady state conditions exist in the aquifer.

In response to temporary climatic fluctuations, there is an episodic pressure fluctuation developed in the regional groundwater flow system during which, under high hydraulic head, water migrates from the aquifer to the surface where it either evaporates or evapotranspires, and allows high salt concentrations to build up in both the soils and the shallow groundwaters. Under lower aquifer pressures, downward directed vertical hydraulic gradients permit the leaching of salts from the vadose zone into the regional subsurface drain - the lead. It is therefore suggested that there is an influx of salts superimposed upon the regional flow system along its length, starting from a point in the vicinity of the upstream limits of the recharge-discharge hinge line, whenever lower hydraulic head conditions prevail in the aquifer and the discharge zone shrinks.

While much of the salt entering the groundwater system during times of lower aquifer pressures could have come initially from the Calivil Formation aquifer, it is likely that part of the recharging salt was deposited on the plain by the stream systems during phases of sheet flooding (cf. Macumber, 1969). A process of episodic salt infusions would explain the observation that the initial increase in the rate of rise of groundwater salinity commences in the Pomppapriel district, that is, in the vicinity of the present-day hinge line.

Since the addition of salts from this source only commences downbasin of the former hinge line, then the most northerly waters, having spent the longest times in the zone of periodic salt accretion, will be the most saline, and, to this extent, residence time is relevant to salinity increase. The rate of salinity increase with distance downbasin will then depend upon such factors as the downbasin flow velocity and the period for which the low-pressure conditions prevail. Of the factors previously listed, the non-steady state flow model of salinity increase best fits into the evaporation-evapotranspiration category, with the rider that the evaporative process takes place at some distance above the aquifer, at or very close to the surface.

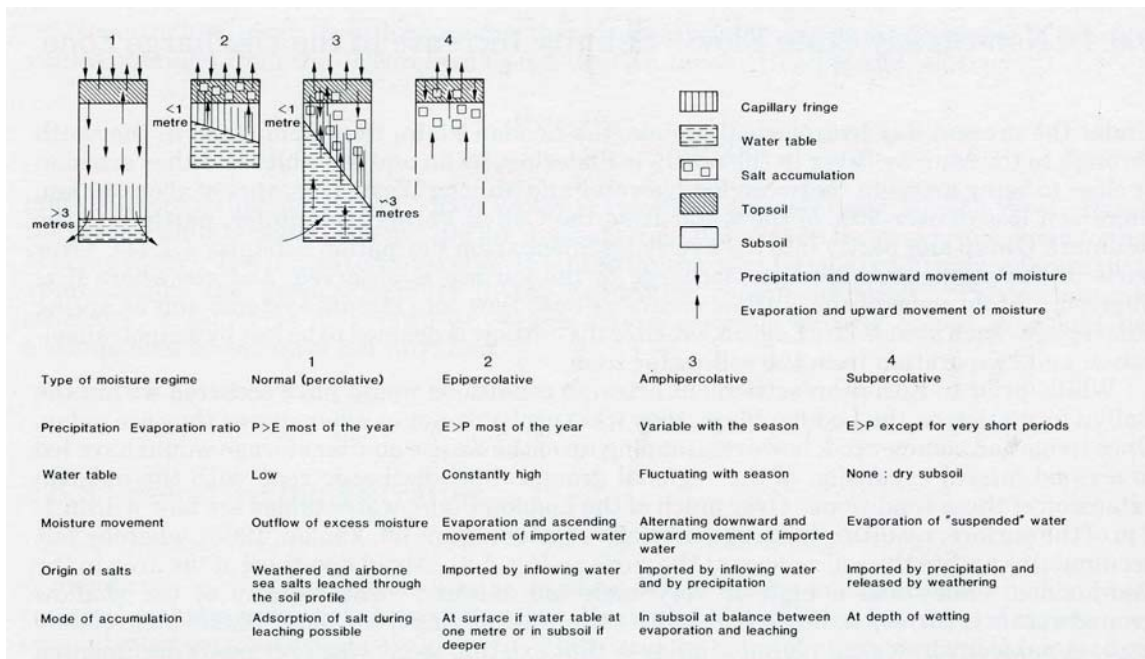


Figure 5.8 - Schematic diagram of characteristic soil moisture regimes, showing their relation to salt accumulation (after Yaalon, 1963).

6. GROUNDWATER DISCHARGE IN THE MALLEE REGION OF NORTHWESTERN VICTORIA

6.1 Introduction

The Murray Basin of southeastern Australia maybe divided into two parts: the Riverine Plain to the east; and the Mallee to the west, with its thin aeolian cover overlying the marine Parilla Sand. The Parilla Sand has had a major influence on the physiography of the Mallee. Hills (1939) first described the north-south ridges, which were later recognized by Blackburn (1962) as being strandlines, formed behind the retreating Tertiary sea (Figure 6.1). The lakes and drainage systems of the Mallee formed in the interr ridge corridors. Erosion of the Parilla Sand has gradually led to the development of the extensive dunefields of the Big Desert, the Little Desert and the Sunset Country (Lawrence, 1975)). However, perhaps one of the most significant influences of the Parilla Sand on the geomorphology of northwestern Victoria arises from its influence as a largely unconfined regional aquifer throughout the entire northwest of the state (Macumber, 1980, 1983).

6.2 Regional Groundwater Discharge Zones of Northwestern Victoria

The Parilla Sand forms a sheetlike layer across the Mallee, averaging about 60 m to 70 m in thickness and having a horizontal hydraulic conductivity ranging from about 0.1 to 10 m/day. It is an unconfined to semiunconfined aquifer, becoming semiconfined only where it is overlain by a thicker development of the lacustrine Plio-Pleistocene Blanchetown Clay, infilling the interridge corridors. The Parilla Sand aquifer is contiguous with the fluvial Calivil Formation (Macumber, 1973); however, being unconfined, it receives water not only from recharge areas upbasin beyond the Mallee, but also from within the Mallee. In general the water is saline, normally ranging from about 2.5 to 4% total dissolved solids.

The Murray Basin is a closed sedimentary basin and a closed groundwater basin. Regional groundwater flow passes downbasin from recharge areas within the highlands and in the Mallee, toward zones of regional groundwater discharge developed toward the centre of the basin. The main Mallee aquifers are the unconfined Parilla Sand, the marine Duddo Limestone and the continental Renmark Group. The main discharge areas are found in the west central Mallee of northern Victoria, southern New South Wales and South Australia. In Victoria, the regional groundwater discharge is best developed in the Sunset Country of far northwestern Victoria, producing a complex pattern of discharge features (Figure 6.2) set in an otherwise aeolian landscape (Macumber, 1980). Groundwater discharge takes the form of small salinas scattered throughout the Mallee dunefields and the very much larger groundwater discharge complexes - the boinkas (Macumber, 1980).

While part of the regional groundwater flow is lost to the groundwater discharge basins of the Sunset Country and nearby areas, the bulk of the flow continues downbasin to the Murray River. Indeed, the Murray River marks the virtual centre of the regional discharge zone, and is the largest single groundwater discharge feature. As the eventual outlet for all through-flowing regional groundwater in the Murray Basin, it is also the ultimate repository for all salt moving via the regional groundwater systems. This is reflected in the salinity of the river, which increases dramatically once the system enters the regional discharge zone in north-western Victoria and eastern South Australia (Figure 6.3).

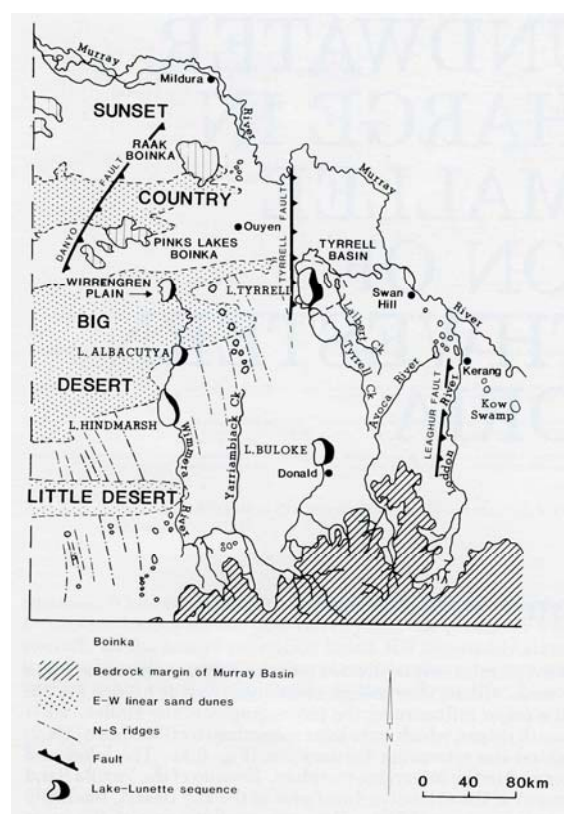


Figure 6.1 - Geomorphology of northwestern Victoria

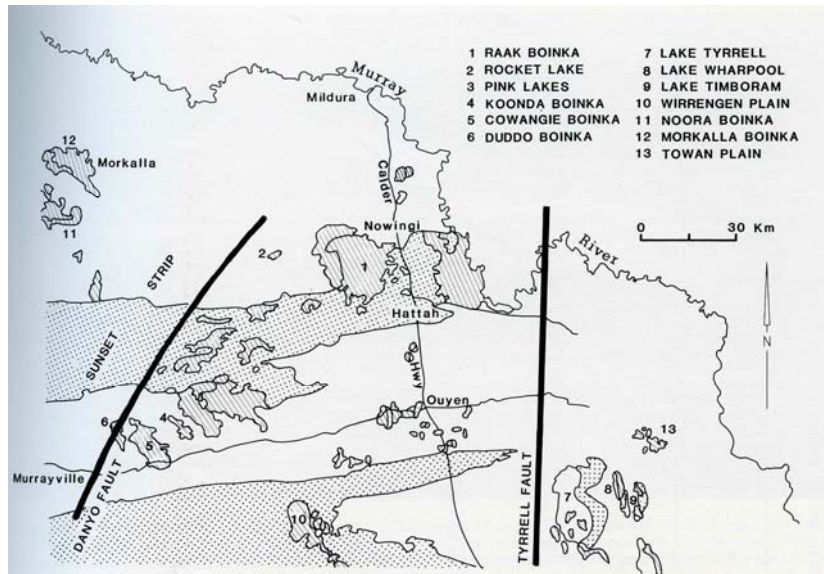


Figure 6.2 - Boinkas and lakes of northwestern Victoria. The Sunset Country lies to the west of the Calder Highway and to the north of the Ouyen - Murrayville Highway.

The Sunset Country is an area where two regional groundwater flow systems merge. From the south comes the west Wimmera system and from the east the main Murray system, feeding the Tyrrell complex on the way. The most detailed information on the flow systems comes from the west Wimmera system where the Parilla Sand is underlain by a major freshwater aquifer, the Duddo Limestone (Lawrence, 1966), which has been extensively utilized throughout the western Mallee (Gloe, 1947; Lawrence, 1975). The Duddo Limestone is separated from the Parilla Sand by an aquitard (the Bookpurnong Beds); however, where information is available in Victoria, it shows that the potentiometric surfaces of both aquifers are roughly at the same elevation (Lawrence, 1975). The potentiometric surface of the Duddo Limestone is, therefore, taken as closely approximating that of the unconfined Parilla Sand.

This is in line with the first appearance of groundwater discharge landforms at the point where the potentiometric surface of the Duddo Limestone rises above the ground surface. In certain areas the potentiometric surface is several metres above ground surface. The water table associated with the Parilla Sand aquifer also cuts the surface, and groundwater discharge occurs at low points in the landscape (Figures 6.3, 6.4 and 6.5).

In the Sunset Country, the potentiometric surface of the Parilla Sand has a general northwesterly slope, and the levels of the floors of the active salinas indicate the approximate depth of the regional water table. Thus, the general concordance of salina levels between the Raak boinka and Pink Lakes boinka (e.g., Raak salinas, 32-34 m AHD, with the exception of Salt Lake; Rocket Lake, 35 m; Mud Lake, 35 m; Lake Crosby, 35 m) shows the general position of the potentiometric surface of the Parilla Sand in the eastern Sunset Country (AHD measurements are relative to the Australian Height Datum, which is essentially that of sea level.)

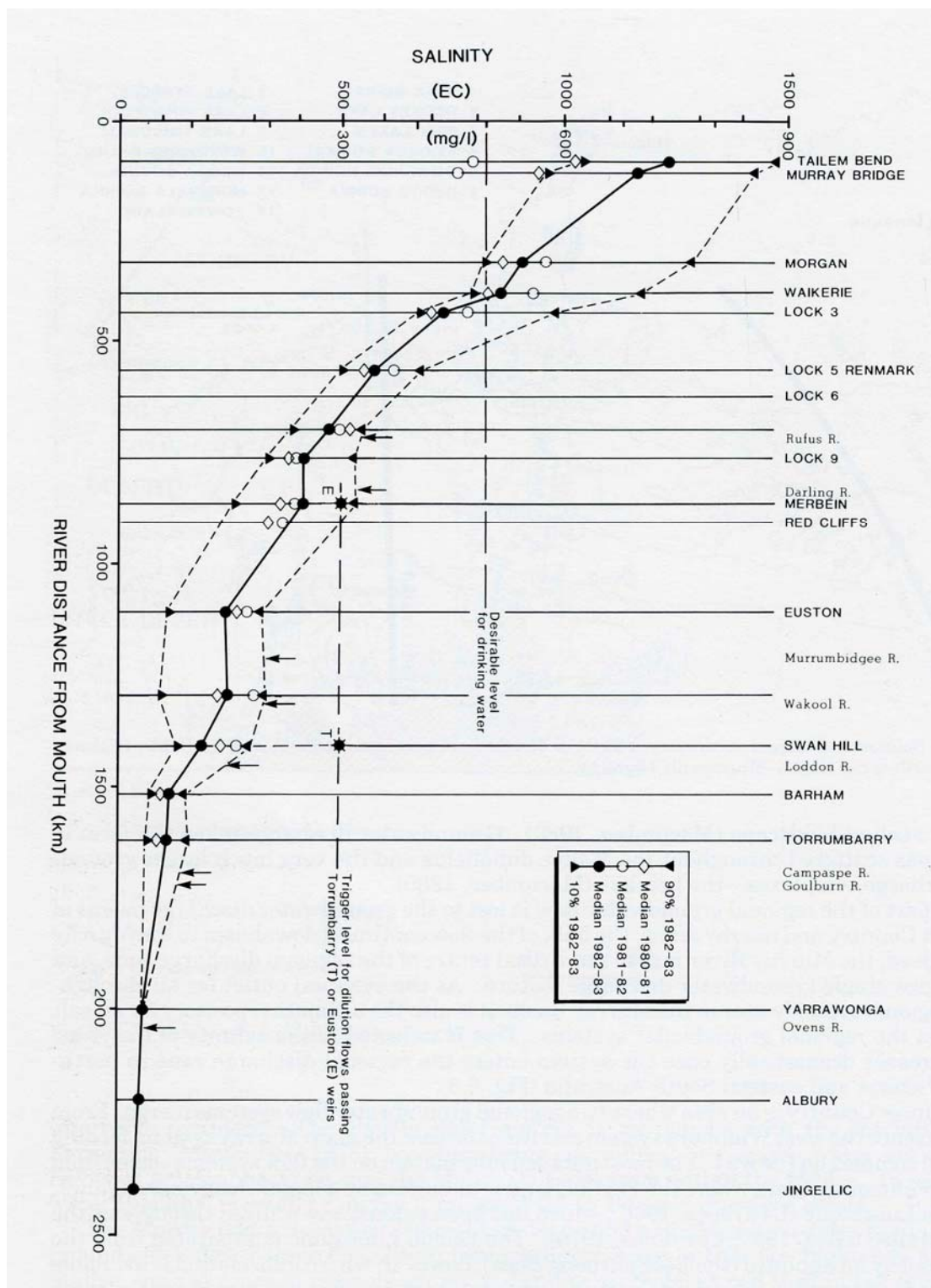


Figure 6.3 - Salinity of the Murray River, showing salt pick up from the groundwater systems as the river passes through the regional discharge zone in northwestern Victoria (River Murray Commission Figure)



Figure 6.4 - Rocket Lake, an isolated groundwater discharge point in the linear dunefield of the Sunset Country. While part of the lake floor now forms an ephemeral salina, the remainder is essentially a gypsum flat with a halophyte cover. Note the truncation of the dunefield at the lake edge, showing that lacustrine activity post-dates the last significant movement of the dunefield at about 15,000 BP.

Groundwater discharge features are also associated with the groundwater flow system, further east; however, in this case, the Duddo Limestone is absent, being replaced by the relatively tight Geera Clay. As a result, groundwater discharge is largely restricted to waters seeping from the Parilla Sand: it occurs wherever the undulating Mallee ground surface intersects the water table of the Parilla Sand aquifer, and its expression is generally either that of a halite salina (ephemerally wet playa lake) or a gypsum flat (with its distinctive halophytic vegetation) (Lawrence, 1975). The area in which regional groundwater discharge processes are active in the Mallee under the present hydrological regime is largely restricted to the regions north of 35°30' latitude and in areas below the 50-metre contour. In Victoria, the four main discharge zones in this category are (Figure 6.2).

- Tyrrell, Timboram and environs
- Raak boinka and minor basins to the northeast
- Murrayville-Underbool district, including the Pink Lakes boinka
- Morkalla-Taparoo area, extending westward to Noora in South Australia

Similar features occur to the north of the Murray River in southwestern New South Wales.

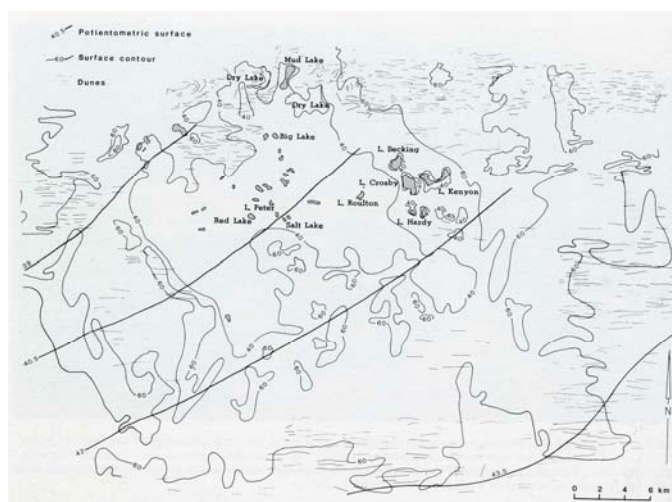


Figure 6.5 - The relationship between topography, potentiometric surface and geomorphology for the Pink Lakes boinka, and the surrounding linear dunefield. While much of the boinka lies below the 40 m AHD contour, the potentiometric surface of the Duddo Limestone is at an elevation of from 39 m to 42 m in this area. (Potentiometric data after Gloe, 1947.)

6.3 Groundwater Discharge Complexes The Boinka

A characteristic of each of the above major discharge zones is the presence of broad, shallow depressions within the Mallee surface. With the exception of the Tyrrell and Timboram lakes, which have developed as terminal systems on the Tyrrell and Lalbert creeks, the discharge complexes show no evidence of fluvial influences. These areas occur as discrete groundwater discharge features, covering, at times, areas of tens to hundreds of square kilometres, and having a very complex geomorphic character. The discharge complexes, of which those at Raak

Plains and the Pink Lakes are the best developed, occupy broad depressions with a distinct outer perimeter (Figure 6.8). The landscape within the depressions is strikingly dissimilar to the surrounding aeolian Mallee landscape and contains its own distinct suite of landforms, which include sand plains, gypsum flats, gypsite (copi) hills, salinas, and various types of source bordering dunes (Figure 6.9). Because of their distinctive nature, the discharge complexes were recognized as landforms in their own right and given the name boinka (Macumber, 1980). Apart from its geomorphic form, the feature which distinguishes the boinka from other landforms is the influence of groundwater discharge as an essential factor controlling the geomorphic processes within the depression. The boinka is a landform complex in which a combination of individual landforms are found: it is not intended that the term should apply to discrete single landforms such as salinas, lake-lunettes or gypsum flats, which may exist on their own or be part of a boinka complex.

Landforms within the boinka may be placed into a general topographic series, ranging downward from sand and gypsum-clay dunes (and islands) to sand plains, clay and limestone plains, gypsum flats and, finally, halite salinas which occur at the lowest points in the landscape.

Groundwater discharge and deflation are the two most active processes in the boinkas, resulting in expansion at the perimeters at the expense of the surrounding linear dunefields. However, in the case of the larger boinkas, it is unlikely that the dunefields originally covered the depressions. Instead, the depressions are seen to reflect an initial tectonic influence. They probably predated the dunefields and, once the latter formed, evolved contemporaneously with them. This is clear in the case of the larger basins such as Raak, where tectonism has strongly influenced the distribution and thickness of the Blanchetown Clay which underlies the Raak boinka.



Figure 6.6 - The Pink Lakes boinka, showing a number of salinas (some with lunettes) scattered across a sand plain (1:80,000).

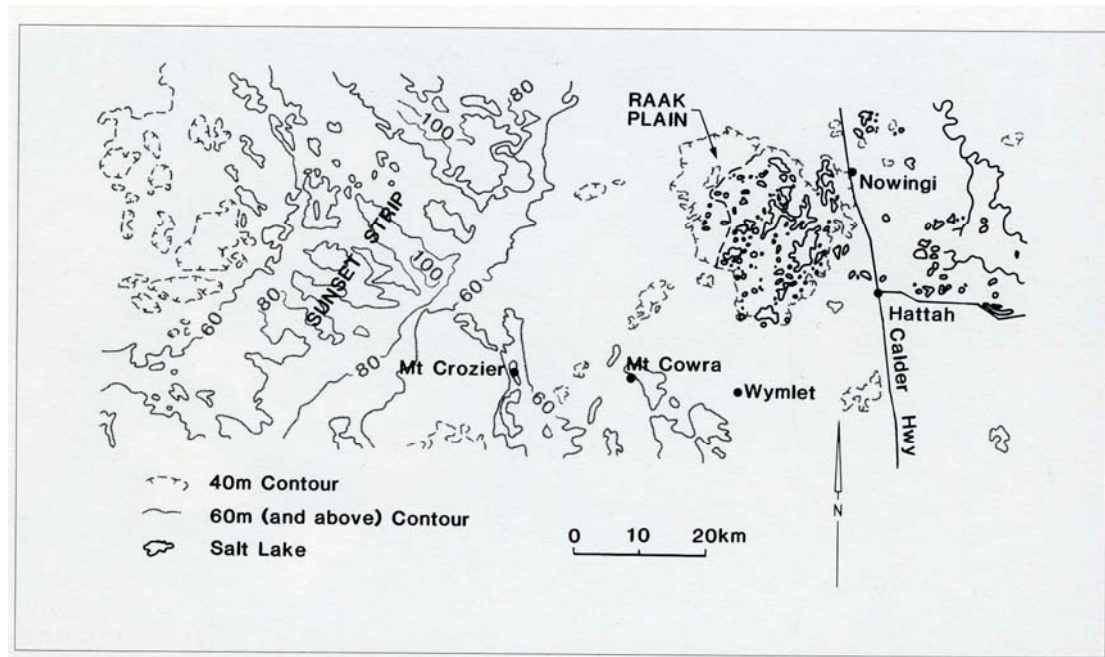


Figure 6.7 - Topography of the central Sunset Country. Groundwater discharge landforms, both active and fossil, lie within the 40 m closed contours on either side of the Sunset Strip, a NE-SW trending horst-like structure on the west side of the Danyo Fault. The fault is marked by a 40 m rise in the topography.

The remnants of NW-SE trending Pliocene strand lines are preserved on the Sunset Strip.



Figure 6.8 - The western edge of the Raak boinka, showing the sharp contrast between the dunefield and the groundwater discharge landscape. The parallel dunes near the pink salina are the Raak 1 and 2 archaeological sites, occupied ca. 7,600 years BP. The salinity of the lake waters is in excess of 250 g/l.



Figure 6.9 - Groundwater discharge landscape in the Raak boinka, showing a number of salinas scattered across a sand plain.



Figure 6.10 - Groundwater discharge from gypsum flats scattered through a subdued linear dunefield near Manangatang. The groundwater comes from the unconfined Parilla Sand aquifer, which has a salinity of 40 g/l.



Figure 6.11 - Contact between the western rim of the Raak boinka and the E-W trending linear dunefield of the Sunset Country. Within the boinka an isolated remnant of the dunefield is flanked on its western side by a series of NE-SW trending copi strand lines, formed by deflation of the former lake floor. Smaller NE-SW trending copi dunes flank gypseous clay islands to the north of the remnant. (1:85,000)



Figure 6.12 - Aeolian islands and sand plain remnants lying within the confines of the main Raak salina at the eastern margin of the Raak boinka. The water is less than 20 cm deep and has a salinity in excess of 300 g/l. It should be noted that there is no surface flow into the Raak boinka, all water being derived from direct precipitation and groundwater outcrop. Annual rainfall is about 250 mm, and evaporation about 2500 mm.



Figure 6.13 - Gypsum flat (detail) on the Raak Plain, at the western edge of the Raak boinka. Groundwater outcrops in small depressions which lie within the arc of a gypseous clay dune about 1.5 km long.



Figure 6.14 - Transverse copi (gypsum) dune in the Duddo boinka, derived by deflation of a former saline area, which is now a grassy flat and largely inactive towards groundwater discharge. A lowering of the water table in the western boinkas is suggested, while the eastern boinkas remain active discharge zones.



Figure 6.15 - Freshwater limestones (represented here by intraformational breccias) occur in the Cowangie and Duddo boinkas. The limestones indicate an earlier freshwater phase within the now saline Cowangie boinka.

6.4 Structural Control of Regional Discharge Features

The distribution of the large groundwater discharge systems is strongly influenced by structural factors, for instance, the Tyrrell Basin, occupied by the Tyrrell-Wahpool-Timboram lake complex, lies immediately upbasin of the Tyrrell Fault (Macumber, 1983), which has since been shown by drilling to extend northward into New South Wales, disrupting the course of the Murray River.

Similarly, it has been shown (Macumber, 1980) that boinkas lying between Murrayville and Underbool lie immediately to the east of the downwarped side of the northeast trending Danyo Fault, which was first recognized by Chapman (1916) and later shown by Gloe (1947) to be a monoclinial flexure passing through Murrayville. Although the fault was only mapped in the Murrayville district where borehole data was available a probable continuation of the uplifted structure lying to the west of the fault can be traced from contours as a topographic high, passing in a north-northeast direction across the Sunset Country, toward Tarrango (Figure 6.7). The rectangular horstlike structure, termed the Sunset Strip (Macumber, 1980), rises 40-60 m above the surrounding, undulating Mallee surface. It has a clear asymmetry, with a significantly steeper gradient on the southeast edge and a more gentle slope to the northwest suggesting a possible tilt block. Remnants of the late Tertiary strandlines showing a northwest - southeast orientation, stand out. In the lower country on either side of the Sunset Strip, the strandlines are largely obliterated and, instead, are replaced by past and present groundwater discharge landforms, which are scattered throughout the dunefields. Apart from the Tyrrell complex all major discharge basins of northwestern Victoria lie adjacent to the Sunset Strip.

The presence of major north-south trending faults lying across the westerly trending flow paths of regional groundwater systems, significantly disrupts the groundwater flow by decreasing the thickness of the aquifer(s) and, therefore, their carrying capacity. This is a crucial factor in the development of regional discharge zones in the Murray Basin. Faults which clearly have this effect are the Danyo Fault, the Tyrrell Fault and the Leaghur Fault.

It is notable that the area to the northwest of the Sunset Strip, extending into South Australia, is shown by Firman (1965, 1973) to have been occupied during Pliocene to Pleistocene times by Lake Bungunnia. The ancient lake was bounded on the south by a structural high called the Pinnaroo Block. It is likely that the Sunset Strip played a similar role to that of the Pinnaroo Block, forming a western boundary to an eastern embayment of Lake Bungunnia, which covered the downwarped topography with lacustrine sediments of the Blanchetown Clay.

6.5 Boinkas of the Eastern Sunset Country

The Raak boinka is the largest single area of active groundwater discharge, and the most striking example of a discharge landscape in Victoria, and perhaps in southeastern Australia (Figure 6.11). The floor is some 15 to 20 m below the level of the surrounding Mallee dunefields, which impinge on the Raak boinka from the west as individual fingers of linear dunes and broad lobes of subparabolic dunes.

The groundwater discharge status of the boinka is clearly seen in the standing water levels of the associated bores. For instance, the Mournpoul 10001 bore, drilled on the shoreline of Spectacle Lake just beyond the eastern limits of the Raak boinka, taps an interval some 40 m below ground surface and has a standing water level about 2 m above the lake floor. The Walpamunda 1 bore on the shore of the main Raak salina is also a flowing bore, as are the Goonegul 10001 and Bitterang 10001 bores from the central areas of the Raak boinka (Figure 10.10). While the salinities of the first two bores are those of the inflowing regional waters, the latter bores have salinities of about 100,000 mg/l. Regional water tables to the immediate northeast of the Raak boinka, in the area between Nangiloc and Colignan, have elevations of about 37 to 40 m AHD, while the average level of the floors of the salinas within the boinka is 33 to 35 m.

The boinka rim is well defined on its western margin by the abrupt change from dunefields to gypsum flats and salinas (Figures 6.7 and 6.10) but is less marked to the east where aeolian sands derived from within the boinka merge with the linear dunefields. The floor is essentially an undulating sand plain across which are scattered scores of salinas and extensive areas of gypsum flat. In the east, there are large areas of interconnected salinas (the main Raak salina, Figure 6.12) dotted with sand and gypseous clay islands. These areas are flanked on their lee side by a north-south running discontinuous transverse dune, initially derived from the salina complex, but now set well back from the shoreline. Classical clay lunettes are uncommon, and instead, clay dunes where present, have often developed as low irregular mounds or ridges along the salina shoreline, or as islands within the salina.

The western quarter of the boinka is an extensive gypsum flat - the Raak Plain (Figure 6.12), where groundwater is removed by means of halophytes. Crystalline gypsum forms in situ, within the sediments, and is mined for the manufacture of plaster of paris. Within the Raak Plain, but rising above it, are a number of islands formed from parts of the linear dunefield, which were cut off from the Mallee during expansion of the discharge zone. On one such remnant (Figure 6.11), a series of arcuate, strandlike, gypsite, copi ridges has prograded westward for about one kilometre across the gypsum flats. Small salinas are found toward the western limit of Raak Plain, adjacent to the basin rim.

While the Raak boinka is seen as being essentially an active zone of groundwater discharge, the more southerly boinkas, developed on the downthrown side of the Danyo Fault between Murrayville and Underbool, show greater variability of their present-day groundwater discharge status. The most active zone occurs in the east - the Pink Lakes boinka, where numerous salinas and gypsum flats are scattered through a sand plain landscape (Figure 6.6). The Pink Lakes boinka is, in many respects very similar to the Raak boinka and, like Raak, lies within the 40 m AHD contour (Figure 6.5). As in the case of the Raak boinka, on passing westward the sand plain-salina component declines, and instead, an essentially gypsum flat-copi rise landscape emerges. Further west again, and separated by northwest - southeast trending ridges, are three similar boinkas all showing extensive development of gypsum flats and north-northeast - south-southwest oriented transverse gypsite-copi rises. The areas have been described in detail by Abbott (1914), who prepared maps and sections showing the distribution of the copi deposits. As is the case elsewhere in the Mallee, the copi often overlies layers of aeolian, crystalline, millet seed gypsum and has been regarded as a weathering product of the crystalline gypsum (Jack, 1921).

The copi rises were formed during previous periods of aeolian activity and are today covered by Mallee vegetation (Figure 6.14). It is notable that the general orientation (NNE-SSW) of the copi rises and the preferential blanketing by copi of the northwestern portion of the Mallee rises point to a northerly component in the predominantly westerly winds during the deflation

of the boinka floors. In many instances, smaller versions of the elongate transverse strands (as found at Raak) occur on the flanks of the rises. The southern boinkas differ from the Raak boinka in having flats developed on freshwater limestone deposits (Abbott, 1914). These border the southern margins of the gypsum flats at Duddo and at Cowangie (Figure 6.15). The limestones indicate an earlier phase of lower salinities within the boinkas: the groundwater is of considerably lower salinity beneath the limestone flats, and it is possible that parallel precipitation of limestone and gypsum may have occurred previously, in a manner similar to that described from Death Valley, California (Hunt et al., 1966). An analysis of the limestone from the Cowangie boinka (Abbott, 1914) is

CaCO ₃	61.87%
Insoluble.	26.02%
H ₂ O	5.54%
MgCO ₃	33.46%
Al ₂ O ₃	2.74%
CaSO ₄	0.37%

The large boinka at Cowangie is an active discharge zone consisting of phreatophyte covered gypsum flats dotted with copi islands. The two smaller boinkas, at Duddo and at Tutye, have saltbush in significantly smaller proportions of their floors and appear to be relics of previous periods of higher water tables and groundwater discharge. In the Duddo boinka, transverse copi dunes border former deflation flats, which are at present well grassed and lightly timbered. The water table is well down, and at Duddo Wells (a shallow well in lacustrine limestone in the south of the boinka) the water table is some 6 m below the grassy limestone flat.

6.6 Boinka Expansion

Whilst tectonics are seen as the temporally distant initiator of the boinkas, deflation is the more immediate cause of ongoing deepening of the basin. However, as in lunette formation, this occurs only under a favourable hydrologic regime. In general, for playa deflation the saline water table must be sufficiently high to allow capillary salinization to discourage growth of vegetation, yet not too high for year-round groundwater outcrop to establish a deflation resistant moist salt crust. Under these conditions any incipient salt crust that may form is dissolved and washed down to the water table during wet periods when the playa has a thin water cover and a temporary recharge status. Once dry, capillary action again proceeds to prepare the playa surface for deflation.

The role of deflation is clearly seen at Lake Tyrrell and the Raak boinka, where there has been a lowering of the lake floor by between 0.7 and 1 m since about 20,000 years BP. Remnants of lake floor sediment are now found preserved at levels above that of the present-day lake floor, as pedestals underlying aeolian deposits deflated during the last major phase of aeolian activity, dated at about 20,000 to 15,000 years BP (Bowler, 1976). The loss of sediment from the basins has left at best only a thin veneer of post middle Pleistocene deposits in the depressions. In many instances, the basin floor is cut directly into either the Parilla Sand or the Plio-Pleistocene Blanchetown Clay. Even in the Lake Tyrrell depression, where sediment may be brought in by the Tyrrell Creek, only about 5 m of laminated gypseous sediment has accumulated on the lake floor over the last 700,000 years (Chapter 2.10). Given a loss of 0.7 m in the last 20,000 years, it is likely that the remaining 5 m represents only a minor part of the material actually deposited on the lake floor and since removed by deflation.

Perhaps the best examples of the ongoing deepening of the basins comes from the Pink Lakes boinka, where ancient lake systems, similar to those existing today, are preserved as sedimentary sequences stranded in the cliffs of the contemporary lakes. At Lake Crosby, for instance, there are some 2.8 m of lacustrine sediments preserved in the western cliff section, the base of the sequence commencing about 2 m above the present lake floor. The net lowering of the lake floor after the cessation of palaeolake activity was about 5 m. The palaeolake is itself incised into an earlier lacustrine sequence of clays and limestones tentatively correlated with the Blanchetown Clay.

Details of the cliff section on the western side of Lake Crosby are as follows:-

DEPTH (M)	LITHOLOGY
0.00- 3.00	Aeolian sand (Recent)
3.00- 3.45	Copi (ancient Lake Crosby sediments)
3.45- 3.80	Laminated gypsum and clay band
3.80- 4.00	Dull reddish brown clay
4.00- 4.22	Copi (? faint laminations)
4.22- 4.30	Dull reddish brown clay
4.30- 4.55	Copi with faint laminations
4.55- 4.75	Grey dense clay
4.75- 5.30	Laminated gypsum and clay bands
5.30- 5.52	Grey dense clay
5.52- 5.77	Light grey, very fine sand and silt
5.77- 8.00	Yellow-grey and orange mottled sandy clay (Blanchetown Clay equivalents) (base of section is present-day lake level)

The laminated sequences are interpreted as representing partly regular (seasonal?) fluctuations in lake level and salinity, while the gypsite copi is seen as a product of subaerial weathering of gypseous lacustrine materials. The interspersed copri beds with laminated gypseous sediments is, therefore, interpreted as representing the existence of dry lake periods of short to intermediate term duration. While the grey clays, which in places contain large groundwater-deposited gypsum crystals, are clearly lake sediments, it may be that the red-brown clay is of aeolian origin. The silt and fine sand mark the onset of sedimentation; where found at the margins of the palaeolake, similar sediments can be demonstrated to have been deposited in a shoreline environment.

Apart from the presence of former lake sediments at high points in the landscape, the expansionary processes active within the boinka are shown by the blocks of linear dunefield preserved as isolated remnants on the boinka floor. Linear dune remnants are best seen in the Raak and Pink Lakes boinkas. At Raak, an isolated remnant (Figure 6.11) rises some 10 to 15 m above the Raak Plain gypsum flat; it is now about 1 km from the windward dunefield at the western lip of the Raak depression. The linear dune remnant is partly blanketed by aeolian copri and has a series of six well-developed copri ridges prograding across the gypsum flats from its northwestern flank. The prograded series is considered to reflect the erosional event noted elsewhere in the Raak boinka, whereby the salina floors have been lowered by about one metre, leaving remnants of the former lake floor as pedestals covered with deflated sediments.

The lateral expansion of the active boinkas is readily understood given the processes of groundwater discharge discussed by McBride and Pfannkuch (1975), who showed that groundwater seepage into a lake system is concentrated in a zone along the lake edge. This process is reinforced in the case of certain larger lakes and boinkas of northwestern Victoria by the presence of dense groundwater brines (Chapter 8.1), which underlie the lakes and boinkas (Macumber, 1980). In these instances, less dense regional groundwater comes into contact with significantly denser groundwater reflux brines underlying the lake, with the consequent development of an interceding Ghyben-Herzberg interface between the two groundwater bodies. The interface develops as a result of the density differences between the regional waters and the brines, in a manner similar to that which occurs in coastal situations. As a consequence, the regional groundwaters cannot underflow the lake or boinka and are forced to the surface, emerging as springs along the shoreline (Chapter 12.4). The regional groundwater flow system in the vicinity of the boinka is, therefore, effectively captured by the boinka. This process is discussed in detail in Chapter 12 of this study.

The zone of groundwater outseepage may extend from several metres up to several hundred metres outwards from the basin margin, which is often varyingly cliffed. At other times, the outseepage zone may extend fingerlike protrusions up the interdune corridors bordering the basin, and these areas may then become the focal points for small salinas. Largely as a consequence of concentrated spring activity, the upper limit of the outseepage zone is an area of continuous, albeit very slow, cliff line recession largely brought on by springhead or basal sapping. Where the groundwater discharge zone passes up the interdune corridors there is a concentration of the groundwater flow lines at the heads of the embayments, thereby resulting in accelerated erosion at these points. Should erosion then proceed transverse to the dune direction, it eventually leads to isolation of dune promontories from the dunefield and their incorporation into the boinka complex, as occurs commonly in the Raak boinka.

6.7 Aeolian Activity

Within the boinka, deflation of the salina flats is accompanied by dune building on the leeward margins; however, as shown by Price (1963), the bulk of the material is lost to the deflation basin as dust. Lunettes are relatively uncommon, with the notable exception of the larger lakes within the Pink Lake boinka. Instead, copi mounds, hills and transverse ridges are the common aeolian landforms, occurring scattered across the lake floor, and not only on its lee side, as is the case with lunettes. The irregular distribution of the aeolian forms reflects slight irregularities in the lake floor, thus resulting in differential deflation. Under favourable hydrologic conditions, deflation occurs at numerous points at any one time, and many small copi mounds or islands may be established adjacent to the deflated area. After a small rise forms, it is likely to become vegetated and remain fixed as a focus for further sedimentation. The remoteness of the discharge basins from the established stream systems prevents regular flooding of the lake floor and destruction of the rises (cf. Bowler and Magee, 1978). Elsewhere within the boinka, more uniform deflation of playas gives rise to elongated north-south trending transverse clay and/or sand dunes on their lee side.

Whatever the final form of the dunes, their lithology is essentially determined by the lake floor lithology. This is most clearly seen in the Raak boinka, where the salina floors are varyingly ferruginized; they are partly cut into ferruginized sand and partly into lacustrine clay. Where the lake floor is sand, islands within the salinas are composed of windblown sand resting on pedestals of ferruginized lake floor sediments (Chapter 7.11, Figure 7.20). Clay dunes do not flank the salina and only appear further to the north, where the floor is cut across a clay infill. In these latter instances, the island pedestals are likewise clay (being former lake bed sediments), while the bulk of the islands consist of aeolian gypseous clays.

The most common form of transverse dune in the boinkas is the NE-SW oriented gypsite copi rise, developed on the salina floors. These occur in all boinkas, including the small Towan Plains boinka, east of Lake Tyrrell. They are best seen in the Morkalla boinka in the northwest Sunset Country, and in the arcuate, strandlike ridges prograding from the linear dune remnant on the Raak Plain. The rises generally show a well-defined northeasterly component but vary from almost N-S to NE-SW. Although fairly straight, they may curve around obstacles and occasionally turn suddenly at right angles. Under the present wind regime, a strong northwesterly component exists only in the winter months of July and August (the two wettest months); however, it is unlikely that the copi rises formed in midwinter when drier conditions necessary for deflation are absent. The E-W oriented, linear dunefields have been shown by Bowler (1976) to have formed under Pleistocene wind regimes, having a similar character to the present-day Mallee winter regime. This situation was achieved by a northward shift of the summer anticyclonic system by at least 5 degrees. The NE-SW orientation of the copi rises differs notably from the typical east-west direction of the linear dunes (see Figure 6.11). An explanation is that the NE-SW trend in the copi rises reflects a predominantly summer wind component, while the more typical east-west trend of the adjacent linear dunefield arises from a more general average annual wind direction.

Although less common than other dune forms, large transverse shoreline dunes occur on the margins of many salinas, and true crescentic lunettes are found on some of the larger lakes such as Lake Crosby, Red Lake and Lake Kenyon (Fig 6.5 and 6.6). All such lakes have smoother shorelines, indicating a surface runoff component sufficient to modify the normally irregular shape of the discharge salina and hence develop the typical elliptical shape associated with stream fed lake basins. Where developed, the lake edge dunes are commonly sand, suggesting deflation of sandy beaches under lake-full conditions. Studies at Lake Tyrrell to the east, and at Lake Wirrengren, to the south, show that such conditions existed in the area in late Pleistocene and early Holocene times.

6.8 Hydrological Equilibrium in the Mallee Landscape

Apart from the large boinkas, there are many small salinas and gypsum flats scattered across the Sunset Country. Similar occurrences are found throughout the northern Mallee (Figure 6.10); for instance, between Nyah and Manangatang numerous small grassed basins are interspersed with small salinas and gypsum flats. The basins are flat-bottomed and often have a pronounced rim. Several large salina-gypsum flat complexes exist in this area, including the Towan Plains which can be categorized as a small boinka. The landscape, where developed, is seen as a palaeo-discharge landscape, reflecting previous periods of higher water table. The flat-bottomed basins occur at low points in the landscape and are, therefore, vulnerable to rising regional water tables brought on by clearing of the Mallee eucalypt cover (Macumber, 1978c). Many previously nonsaline, grassed depressions in the Manangatang area have become saline (Chapter 4.1.8) and are covered with a white salt crust in summer. Additional comments on recent changes in hydrologic equilibria are given in Chapter 4. As in the Sunset Country, where the level of the active salinas gives an indication of the level of the water table associated with the Parilla Sand aquifer, the case is similar in the Manangatang area, where the floors of the recently salinized depressions are at an altitude of about 50 m (AHD), (P. Dyson, pers. corn.).

The effect of previous phases of high water tables on the physiography of the Mallee is perhaps best seen in the Sunset Country, where plains within the linear dunefields have a groundwater discharge origin. Larger grassed plains occur to the east of the Sunset Strip, while smaller, more elliptical plains occur to the west of the Strip. The latter basins are of similar size and shape to Rocket Lake (Figure 6.4) and are assumed to have had a similar character during times of active groundwater discharge. The surface levels of the plains are, in many instances, only a few metres above that of the Noora boinka - a large area of gypsum flat, which extends westward from Victoria to Ingalta and Noora in South Australia.

Many grassed depressions show a clear truncation of the linear dunefields on their western margins, and lunettes are occasionally present. Truncation of the linear dunes (cf. Rocket Lake, Figure 6.4) implies the existence of high water tables some time after the last significant phase of dune mobilization—dated by Bowler (1976) as terminating at about 15,000 years bp. Evidence indicating the occurrence of a phase of postglacial high water budgets has come from northern Victoria (Chapter 3.5 - 3.9). Given the absence of stream systems in the Sunset Country, water table rises in the unconfined Parilla Sand are seen as reflecting higher precipitation than occurs today in the northern Mallee.

With the rise in water tables, the size and number of discharge points increase until a new equilibrium between recharge and discharge is established and the water table stabilizes. A similar readjustment is occurring today throughout the Mallee, where the clearing of trees has caused increased infiltration, leading to rises in water tables and the reactivation of previously dry salinas. The large number of recently salinized depressions provide a clear record of the upward trend in water tables throughout the northern Mallee region. This rise echoes similar trends in the regional aquifers of the Riverine Plain and must be seen as a significant threat to the lower areas of the Mallee.

7. HYDROCHEMISTRY OF THE GROUNDWATERS IN THE PARILLA SAND AQUIFER

7.1 General

Within the Mallee, the Parilla Sand aquifer ranges in thickness from 60 m to 100 m. It is unconfined to semi-unconfined and overlain by a thin aeolian blanket. Recharge within the region is essentially by direct infiltration of rainwaters as there are no significant surface streams bringing low salinity recharge waters from outside the regions to influence the chemistry of the through-flowing groundwaters. Within the Mallee, the regional hydraulic gradients range from about 0.2 m/km to 0.3 m/km. The hydraulic conductivity of the Parilla Sand is about 2 m/d, and groundwater flow velocities are very low, being between 0.001 m/d and 0.005 m/d.

7.2 Regional Trends in the Hydrochemistry of the Central Mallee

In order to gain a clear view of the chemical character of the groundwaters within the Parilla Sand aquifer, some sixty shallow and deep bores into the Parilla Sand were sampled from the vicinity of the Tyrrell Basin. Of these bores, twenty were chosen which were sufficiently distant from, and upbasin of, any modifying influences (present or past) of the Tyrrell-Wahpool-Timboram lake complex; these waters were seen as being representative of the regional groundwater within the Parilla Sand aquifer of the central northern Mallee of northwest Victoria (Table 7.1). For comparison, 14 samples of bore waters were collected from the Nangiloc-Colignan area, 100 km downbasin, to the northwest (Table 7.2).

TABLE 7.1 - Regional groundwater from the Parilla Sand aquifer of the Tyrrell Basin

	Seawater*	Parilla water (mg/l)	Std dev	Conc. ratio with respect to seawater	Excess with respect to bromide
Bicarbonate**	140	249		1.78	1.80
Bromide	65	64.2	13	0.99	1.00
Chloride	19,000	19,757	4,118	1.04	1.05
Sulphate	2,690	3,188	834	1.19	1.20
Calcium	400	470	356	1.17	1.18
Magnesium	1,300	1,353	310	1.04	1.05
Sodium	10,500	11,077	2,227	1.05	1.06
Potassium	380	113	44	0.30	0.30
Silica	6	56	23	9.38	9.47

* Analysis from Collins, 1975 ** From non-acid waters only

TABLE 7.2 - Parilla Sand water from Nangiloc-Colignan district* (comparison with seawater)

	Seawater**	Parilla water (mg/l)	Std dev	Conc. ratio with respect to seawater	Excess with respect to bromide
Chloride	19,000	19,802	2,606	1.04	1.00
Sulphate	2,690	3,596	563	1.34	1.28
Calcium	400	412	320	1.03	0.99
Magnesium	1,300	1,484	224	1.14	1.10
Sodium	10,600	11,645	1,763	1.10	1.05
Potassium	380	72	33	0.19	0.18

* These waters were sampled at a very early stage in the study, when silica and bromide analyses were not carried out.

**Analysis from Collins, 1975

The range of salinities of Parilla water is from under 20,000 mg/l to over 50,000 mg/l. A comparison of waters from east of Lake Tyrrell with waters from Nangiloc-Colignan (Table 7.3) confirms the absence of significant downbasin evolutionary trends in the chemistry apart from perhaps the loss of potassium. It also suggests that, with respect to the major ions, the groundwaters have reached a state of chemical maturity beyond which further modification by such processes as ionic exchange has virtually ceased.

TABLE 7.3 - Comparison of Tyrrell groundwater with Nangiloc - Colignan groundwater (mg/l)

	Cl ⁻	SO ₄ ²⁻	Na ⁺	Mg ²⁺	Ca ²⁺	K ⁺
Tyrrell district	19,757	3,188	11,077	1,353	470	113
Nangiloc-Colignan	19,802	3,596	11,645	1,484	412	72

This is not to say that elsewhere within the basin regional trends do not occur in the chemical parameters within the Parilla Sand; indeed, significant increases in salinity occur at the margins of the Parilla Sand outcrop, adjacent to the highland front. This is to be expected, given the more variable forms of recharge operating at the Mallee edge, the decreased aridity, and the presence of zones of mixing with inflowing non-Mallee groundwaters. (See Figure 56a, Lawrence, 1975).

7.3 Parilla Sand Hydrochemistry Origin of Salts

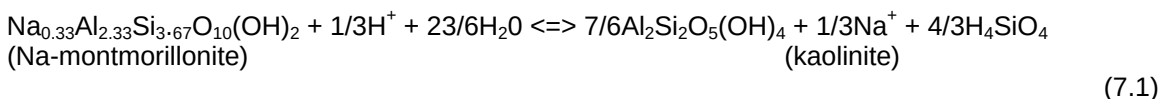
Table 7.1 is a summary of the Parilla Sand hydrochemistry in the Tyrrell region. Here, the average concentration (in mg/l) of each ion is compared with that of seawater. Bromide was chosen as the most useful ion with which to compare other ions. Like chloride, it is relatively inert, but, unlike chloride, it effectively remains in solution, not forming its own minerals under the hypersaline conditions produced by evaporitic concentration in the Mallee playa lakes. This enables further ratio comparisons when examining the ongoing concentration of brines and bitterns into and beyond the halite precipitation phase, which is often reached in the lakes. For comparison with seawater, chloride ratios of the regional Parilla Sand water in the eastern Mallee are given in Table 7.4.

TABLE 7.4 - Ionic ratios of seawater and Parilla water

	Cl/Na	Cl/Br	Cl/SO ₄	Cl/Mg	Cl/Ca	Na/K
Seawater	1.8	292	7.2	14.6	47	28
Parilla water	1.8	308	6.2	14.6	42	98

Perhaps the most obvious feature of the tables is the striking resemblance of the Parilla water to seawater, when the Br⁻, Cl⁻, Ca₂⁺, Mg₂⁺ and Na⁺ ions are considered. This not only applies to the various ionic ratios but also to the total salinity. Table 7.1 shows the average concentration of each ion. The extent to which the Parilla water resembles seawater is demonstrated by comparing the concentrations of the respective ions (the concentration ratio). A further comparison can be made by taking the Br⁻ ratio as an index of concentration, against which other ratios may be compared (excess factor). The regional Parilla Sand water appears to be essentially an oceanic type water, in which the chloride, magnesium, sodium and bromide concentrations are almost identical to the concentrations in seawater; calcium and sulphate are slightly in excess of the concentrations in sea-water; potassium is less; silica, and to a lesser extent bicarbonate, have increased concentrations relative to the other ions.

Low potassium concentrations are a common feature of many groundwaters (Collins, 1975), being readily absorbed on clay minerals. Clay minerals are also likely to be the source of the high silica content. Under normal conditions, the dominant dissolved silica species is H₄SiO₄ (White et al., 1956). Despite the long residence time in the quartzose Parilla Sand aquifer, the solution of quartz is unlikely to have contributed significantly to the silica levels because it has been shown that quartz does not greatly influence the silica content in groundwaters (Freeze and Cherry, 1979). Instead, the main contributions come from the aluminosilicates such as feldspars and clays. There are thin discontinuous clay seams in the Parilla Sand, the main clay unit being the Tragowel Clay Member, which, although best documented further east in the Loddon plain has lateral equivalents in many areas of the Mallee. The production of H₄SiO₄ may occur as a result of many reactions involving the incongruent dissolution of montmorillonitic and illitic clays to produce kaolin. This process is greatly assisted by acidic conditions found within the aquifer, where pH values are commonly less than 4.5 (see Chapter 7.6 and following). Examples of the incongruent dissolution of aluminosilicate minerals have been given in Chapter 5.4, and one of these reactions is repeated below:



A silicate stability diagram (Figure 7.1), based on 23 bores within the study area having a salinity range from 45,000 to 25,000 mg/l (the regional groundwaters), shows that the neutral to slightly alkaline waters are in equilibrium with albite. The equilibrium differs from that of seawater on account of the higher H₄SiO₄ content. With increased acidity (lower Na⁺/H⁺), the equilibrium

moves obliquely downward, firstly, into the Na–montmorillonite field, and then into the kaolinite field where virtually all the acid groundwaters lie. At the same time as the H^+ ion increases, there is a gradual increase in the H_4SiO_4 content of the groundwaters.

The production of H_4SiO_4 by the incongruent dissolution of aluminosilicates requires the consumption of H^+ ions, with a progressive increase in the pH of the waters, for example, equations 7.1, 5.5 and 5.6. However, for the Parilla Sand water this is not the case, since both H^+ and H_4SiO_4 increase together. An explanation is that the acidity is not derived from the aluminosilicate dissolution process but, instead, results from other hydrochemical processes such as oxidation–reduction reactions. If a separate source of H^+ ions is present, then chemical equilibria will be driven toward water that is in equilibrium with kaolinite, with the release of H_4SiO_4 and cations into solution.

Since the regional water has an essentially oceanic character, but with the addition of calcium, sulphate, bicarbonate and silica, and less potassium, it is necessary to consider its origin. Oceanic characteristics have been widely reported in southern Australian surface waters and groundwater (see below), and an oceanic source has been used to explain the high chloride content of Australian waters. Indeed, the question of the origin of the high chloride content of the salts in the streams, soils and groundwaters of Australia has been a contentious issue for many decades.

It is not the aim of this work to discuss at length the genesis of the regional Parilla Sand water. The main intention is to accept these waters as a starting point, and to examine their evolution in the light of processes active in the groundwater discharge landscape and semiarid environment. However, given their oceanic character, which is perhaps more evident in this study than in any similar study so far carried out in southeastern Australia (see Lawrence, 1975; Maddocks, 1966), some comment is necessary on the question of connate waters, cyclic salts, or rock weathering salts.

7.3.1 Rock Weathering

The oceanic character of the regional Parilla Sand water is undeniable, and in the author's opinion (Macumber, 1983) it cannot be explained by any simple rock weathering process alone. This is brought out by the high chloride and bromide contents of the waters - both ions being only minor elements in the earth's crust. Mason (1966) gives the following average amounts in the crust: chloride 130 ppm, and bromide 2.5 ppm. By comparison, sodium is 28,300 ppm, magnesium 20,900 ppm, potassium 25,900 ppm, and iron 50,000 ppm. Collins (1975) gives values of 180 ppm chloride in shales, 10 ppm in sandstones, and 150 ppm in carbonates; for bromide, the respective values are 4, 1, and 6 ppm. For granites, Mason (1966) gives values of 50 ppm for chloride and 0.5 ppm for bromide. Finer grained shales may retain their original connate waters for some considerable time and may be reinfused with chloride-rich solutions if resubmerged beneath the sea. Therefore, a higher value for chloride (1,466 ppm) is given by Billings and Williams (1967) for deeply buried shales: they consider that the lower values are representative of shales that have been largely leached of chloride. With the foregoing in mind, the following possible explanations for the similarity of these waters to oceanic waters are canvassed:

1. Original connate seawater still retained in the aquifer (cf. Lawrence, 1975).
2. Cyclic salts of oceanic origin contained in rainwaters falling within the Mallee and Mallee catchments (cf. Anderson, 1941)
3. Oceanic character - the end result of a natural metamorphosis of mixed waters containing cyclic salts and rock-derived salts released during weathering within the highlands (cf. Chebotarev, 1955).

7.3.2 Connate Water

Lawrence (1975), commenting on groundwater in the Parilla Sand, shows that the oceanic number (Anderson, 1941) increases toward Lake Tyrrell, where it reaches 100. A connate water source is then assumed. The oceanic number is based on the observation that in seawater one equivalent of chloride radical is associated with 0.107 equivalents of other negative radicals. Lawrence (1975) considers that the oceanic character of the Parilla Sand water stems from the incomplete flushing of the aquifer, with additional connate water rising from deeper aquifers under favourable hydraulic heads. Clearly, from the data in Tables 7.1 and 7.4 there can be no mistaking the chemical affinities of the Parilla Sand water with those of seawater; however, this alone does not provide absolute proof of a connate seawater source, since as is shown earlier (Chapter 5.3) the regional groundwaters in the Calivil Formation aquifer feeding the Parilla Sand aquifer have already attained an oceanic character long before passing into the marine sequences. If it is assumed that the Parilla Sand waters are connate, then other implications immediately arise, for instance, that there has been no significant flushing of the Parilla Sand aquifer since the early Pliocene regression—this, despite all the potentiometric evidence showing a general northwesterly direction for groundwater flow in northern Victoria (Figures 41, 42—Lawrence, 1975). Visual evidence of ongoing lateral ground water movement is seen by the fossil discharge landforms preserved in the Mallee landscapes and in the present-day movement of saline waters into and out of the large discharge lakes and boinkas (Macumber, 1980) and into the Murray River. While the permeability of the regional unconfined Parilla Sand is low, as are the groundwater flow rates within the aquifer, it is considered that there has been more than sufficient time for general throughflow and flushing of original connate waters in the Parilla Sand. The other suggestion of Lawrence (1975) is that upward migrating connate waters may be contributing to the salinity of the Parilla Sand aquifer. In order to determine the extent of upward migration of water from the deeper aquifers in the Tyrrell district, two bores were drilled to Palaeozoic basement at about 290 m. The bores proved a simple three-layer system, with a 70 m thick Parilla Sand unit being underlain by a 170 m thick clay and silty clay sequence—the Geera Clay (Lawrence, 1966). This was in turn underlain by the lower Tertiary Renmark Group, consisting of interbedded sand and ligneous clay. Of these three units, only the Parilla Sand and the Renmark Group are meaningful aquifers, while the Geera Clay forms a thick clay aquiclude with a very low hydraulic conductivity: $K_x = 4 \times 10^{-5}$ m/d; $K_v = 2 \times 10^{-5}$ m/d, (Lawrence, 1975).

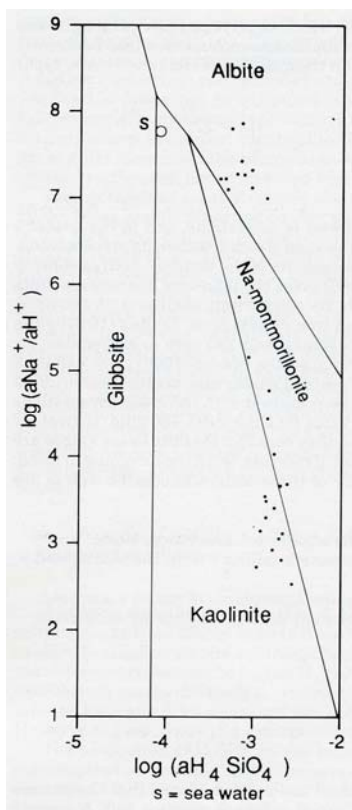


Figure 7.1 - Silicate stability diagram—regional Parilla Sand groundwater
s = seawater

Within the Tyrrell Basin, the Geera Clay aquiclude is sufficiently tight to prevent any significant movement of water from the deeper aquifer. Closer to the lakes, where the Parilla Sand is filled with high density reflux brines the vertical hydraulic gradient is shown to be from the Parilla Sand down into the Geera Clay (Chapter 12).

Stable isotope data (D/H) measurements give values well within the range for waters of atmospheric origin, whatever the salt origin.

7.3.3 Atmospheric Salts

A considerable amount has been written about the possible source of oceanic salts in rainwater ever since the idea was first suggested by Pospeny in 1877 to explain the similarity in chemical composition between many inland lakes and seawater. Salt nuclei in the atmosphere are derived from the bursting of small foam bubbles, thereby producing small water droplets which are swept up into the atmosphere (Woodcock, 1953). Evaporation of the droplets produces a small salt particle which may become the locus for raindrop concentration. Twomey (1955) showed that southwesterly air streams passing across southeastern Australia contained sea salt particles, which were removed in atmospheric precipitation. Anderson (1941) had earlier analyzed rainwater for salt content and, as a consequence, attributed salts in the O'Shannassy River catchment of southern Victoria and the lakes of southwestern Victoria to 'cyclic salts'. Hutton and Leslie (1958) did further rainwater analyses and, using the recorded average rainfalls, calculated the average salt fallout over much of Victoria. Douglas (1968) related the composition of streams in several catchments of eastern Australia to rainfall, while Maddocks (1967), in a detailed study of 52 lakes and rivers of southwestern Victoria, found fairly constant chloride to bromide and chloride to sodium ratios, which were similar to those of seawater. Quoting the conclusion of Hutton and Leslie (1958) that 67 kg of sodium chloride is added to each hectare per year in the Western District, Maddocks concludes that the principal source of salts in the waters he examined was from the atmosphere, as suggested by Anderson.

Detailed studies of the regional distribution of atmospheric salts have been carried out for many parts of the world, e.g. Eriksson (1955) in Scandinavia, Sugawarra (1967) and Kombayasi (1964) in Japan, and Junge and Werby (1958) in the U.S.A. This work has shown that salt content and ionic ratios alter on passing downwind from the source. A decrease in the total dissolved solids content with distance from the coast was noted by Hutton and Leslie (1958) and by Junge and Werby (1958). Loewengart (1961) noted that the chloride content of rainwater decreased exponentially with distance from the ocean; he considered that airborne salts were the major source of salinity in Israeli water.

Junge and Werby (1958) note that, while the major source of atmospheric chloride is the ocean, the Cl/Na ratio falls away rapidly with distance from the coast in response to the addition of terrestrial sodium. Increased Na/Cl ratios were observed in Japanese studies, which show catchment enrichment in calcium, magnesium and sulphate, relative to both chloride and sodium in rainwaters. Hutton and Leslie (1958) also recorded a fall in chloride content with distance from the sea and a consequent increase in Na/Cl ratios.

7.4 Unconfined Aquifers as a Repository for Atmospheric Salts

That large quantities of salts have been added to the land surface and to streams and lakes of Victoria is evident from the work of Anderson (1941) and Hutton and Leslie (1958). The detailed studies of streams, lakes and catchments cited above demonstrate an oceanic character of surface waters based largely on the presence of oceanic ratios for such ions as chloride, sodium and bromide.

This oceanic character is also evident in the groundwater systems. Indeed, since the transit time for unconfined aquifers may be tens or even hundreds of thousands of years, it is clear that such aquifers are salt sinks and contain a long record of salt accessions from atmospheric sources. The aquifer chemistry provides an insight into the nature of the incoming salts and their modification in both the vadose and phreatic zones. In these cases, the groundwaters should not only show the classic downbasin evolutionary trends of sulphate reduction, ion exchange, mineral precipitation, ultrafiltration etc., but will also have imposed characteristics of the recharging waters. The problem then becomes how to disentangle the downbasin evolutionary trends from any possible atmospheric salt overprint.

7.5 Oceanic Character of Surface Waters and Groundwaters

In the case of the Parilla Sand aquifer, any significant trends in hydrochemistry, including overprint from atmospheric salts, have already occurred prior to their reaching the Tyrrell Basin. This is shown by the similarities between groundwaters at Lake Tyrrell and those from Nangiloc–Colignan, a further 100 km downbasin. Evolutionary trends must therefore be examined closer to inflow areas at the basin margins, where the groundwaters undergo their initial salinity increase. A number of trends were observed in the Parilla Sand by Herczeg (1977), who, using the published data from Lawrence (1975) and a further 26 samples collected from western Victoria, noted that except in the very dilute samples, the waters are near equilibrium with respect to aqueous species and clays. Figures produced by Herczeg (1977) show that, except in the case of very dilute waters ($\text{Cl}^- < 200 \text{ mg/l}$), the Mg/Cl ratio and the $(\text{Na} + \text{K})/\text{Cl}$ ratio remain approximately constant and similar to seawater despite increasing salinities (chlorinities ranging from 200 mg/l up to 10,000 mg/l). Herczeg's results show a similar trend to that which occurs in the Calivil Formation in the Loddon Valley (Chapter 5.3).

On the basis of NaCl and Cl/Br ratios, Maddocks (1967) suggested that the salts in streams and lakes of western Victoria are atmospheric salts of oceanic origin. Both Maddocks and Herczeg note a loss of magnesium in the lower salinity range. Maddocks suggests that it has been precipitated as dolomite in a lacustrine setting, while Herczeg attributes the loss to cation exchange. However, in each case the magnesium proportion becomes constant once the salinity rises. Maddocks's magnesium values have been used to produce Table 7.5 below:

TABLE 7.5 - Average magnesium percentage (by weight) of southwestern Victorian lakes and maars (after Maddocks, 1967).

Salinity range (mg/l)	No. of samples	Average Mg (% by weight)	Std dev
< 5,000	13	15.8	5.0
10,000 - 20,000	9	10.2	3.2
20,000 - 50,000	6	10.2	1.6
50,000 - 100,000	5	9.3	1.8
> 100,000	3	6.4	3.5

The extent to which the magnesium content of the lake and river waters of southwestern Victoria approximates that of seawater can be demonstrated by use of the Mg/Br ratios. In the case of seawater, the Mg/Br ratio is 20. A least square fit analysis of the 45 lake waters analyzed by Maddocks gives the relationship as

$$\text{Br}^- = 0.0574 \text{ Mg}^{2+} + 2.43 \quad (7.2)$$

The correlation coefficient for this relationship is 0.919.

For lakes and rivers taken together, the relationship is

$$\text{Br}^- = 0.0577 \text{ Mg}^{2+} + 1.44 \quad (7.3)$$

with a correlation coefficient of 0.992

This gives a Mg/Br ratio of 17.3 to 17.4 for the surface waters of southwestern Victoria, compared to the oceanic ratio of 20. A similar relationship can be shown for the lakewaters and groundwaters of northwestern Victoria. In figure 7.2, data are plotted from about 200 samples of lakewater and groundwater from the Tyrrell Basin: salinities range from 20,000 mg/l to 330,000 mg/l. As in the case of the brackish to saline waters of southwestern Victoria, there is no indication of any significant loss of magnesium over this salinity range.

On the evidence of Herczeg (1977) and Maddocks (1967), it appears that certain oceanic characteristics are already present in the groundwaters and surface waters of western Victoria while still at an early stage of their evolution. This relationship was borne out by the data from the unconfined Parilla Sand aquifer of the Mallee Region of northwestern Victoria. To further test the extent to which this oceanic character occurred in Victorian waters, a set of samples was obtained from fractured bedrock aquifers developed in weathered Palaeozoic sediments and metasediments in the central Victorian highlands (Table 7.6, samples by courtesy of Dr J.J. Jenkin). For comparison, groundwater from the Parilla Sand aquifer of the Tyrrell Basin and seawater are added.

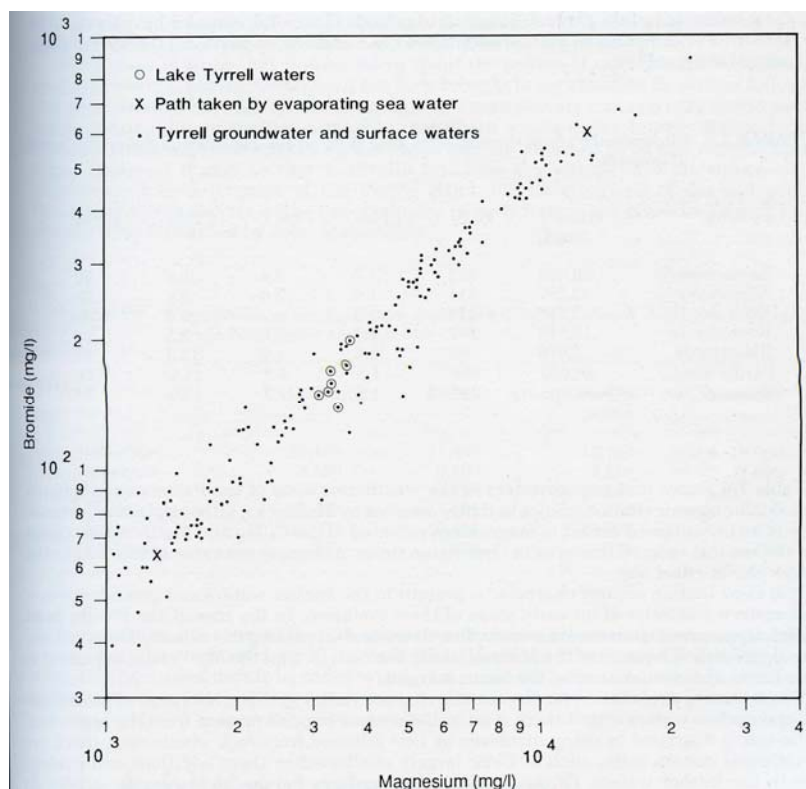


Figure 7.2 - Log bromide vs. log magnesium for all surface waters and groundwaters in the Tyrrell Basin

TABLE 7.6 - Groundwaters from the weathered zone of Palaeozoic bedrock aquifers of the Central Highlands

Locality	Salinity (mg/l)	Cl/Br	Cl/Na	Cl/SO ₄	Cl/Mg	Mg/Br
Kamarooka 5b	22,033	333	1.7	3.4	8.4	39
Kamarooka 5a	23,297	314	1.9	3.0	9.4	33
Knowsley lb	13,717	275	1.9	4.2	9.0	30
Knowsley la	16,849	239	2.0	5.5	9.3	26
Runnymede	2,078	256	1.4	4.0	12.8	21
Parilla Sand	36,000	308	1.8	6.2	14.6	21
Seawater	35,000	292	1.8	7.2	14.6	20

Table 7.6 shows that groundwaters in the weathering zone of the Palaeozoic sediments have similar oceanic characteristics to those observed by Maddocks, although in this instance there is an unexplained surfeit of magnesium reflected in the Cl/Mg and Mg/Br ratios. Given that the sea

last entered this area in Ordovician times, a connate seawater source within the aquifer can be ruled out.

It is clear that an oceanic character is present in the surface waters and groundwaters of southeastern Australia at an early stage of their evolution. In the case of the Parilla Sand aquifer, this oceanic character is even further developed on passing downbasin. Thus, at Lake Tyrrell, towards the centre of the Murray Basin, the Na^+ , Cl^- and the Mg^{2+} ratios are closer to those found in seawater than at the basin margins.

The tentative explanation for the oceanic characteristics in this wide range of groundwaters and surface waters is that there is an initial oceanic imprint present from the beginning. While this is disguised in many instances by ions released from rock weathering, there are nevertheless certain ratios such as Cl/Br largely unaffected by these additions and present even in the fresher waters. Of the available explanations for the high chloride content in streams, soils and groundwaters and for the oceanic Cl/Br ratios in waters of virtually all salinities across a wide range of environments, the most plausible is an atmospheric salt source of initially oceanic origin. If this is the case, then trends in oceanic character will become more apparent in waters further from the influence of rock weathering additions.

The main argument against the above explanation is that, while under the present-day climatic regime the Na/Mg ratios in rainwater remain fairly constant at about that of seawater (Hutton and Leslie, 1958), there is an increase in the Na/Cl ratio with distance from the coast. This is certainly true today, but it can also be asked how relevant are the present-day rainwater analyses to rainfall and recharge events which took place some considerable time in the past and under climatic conditions unlike those of today.

Whatever the initial chemical characteristics of the groundwaters, there is also the downbasin chemical evolution of groundwater along the lines suggested by Chebotarev (1955), who concluded that groundwaters tend to evolve towards the composition of seawater. The development of an oceanic character arises from the combination of an initial oceanic overprint, which is steadily reinforced with distance from the basin margins, coupled with a natural trend toward an oceanic character as a consequence of processes acting within the soil water zone and the aquifer proper. The general pattern of chemical evolution, seen in the waters on the margins of the basin and in the Parilla Sand, conforms to this model. In the case of the Parilla Sand aquifer, the result is not only similar ionic ratios to that of seawater but also a similar ionic concentration, maintained virtually across the basin.

While there is much that is speculative about the origins of salts in surface waters and groundwaters of southern Australia, it has been brought to my attention by various colleagues that groundwaters having a similar composition to seawater are common in semiarid parts of southern Australia, in similar settings to that which occurs in the Murray Basin. In most instances these waters are taken at face value and are assumed to be of a connate marine origin. However, it may be that in certain instances the waters have undergone similar evolutionary trends to those of the Parilla Sand and have arrived at the end point of Chebotarev's evolutionary path. Two examples of such waters are given in Table 7.7 - the samples were forwarded by John Waterhouse.

TABLE 7.7 - Comparison of groundwaters from Loch Coalfield, South Australia with regional Parilla Sand water and seawater

	Lock 1	Lock 2	Average Parilla water	Seawater
Chloride	23,436	17,489	19,757	19,000
Sulphate	3,560	2,700	3,188	2,690
Calcium	600	415	470	400
Magnesium	1,431	1,091	1,353	1,300
Sodium	12,894	10,451	11,077	10,500
Potassium	235	210	113	380

In conclusion, it is stressed that the above considerations are not meant as a comprehensive comment on the origins of salts in the Australian landscape, but instead are put forward *en passant* as additional data to be added to that vast amount of information already accumulated on this subject.

7.6 Acid Groundwaters and Springwaters in Northwestern Victoria

7.6.1 Acid Groundwaters

Acid groundwaters ranging in pH from about 4.5 to less than 3.5 occur extensively in the Parilla Sand aquifer of northwestern Victoria. They are also commonly found in the Palaeozoic bedrock aquifers of central Victoria.

A detailed study of the distribution and chemical affinities of acid groundwaters in the Parilla Sand aquifer and the associated springwaters was made in the Tyrrell Basin of the central Mallee region. Acid regional groundwaters are widely distributed throughout the Tyrrell Basin. The main exception to this is in the case of that part of the Parilla Sand aquifer recharged by the neutral pH Timboram - Wahpool reflux brines and the Tyrrell reflux brines (Figure 7.3). Eastward of the Tyrrell Basin, bores containing acid groundwaters are interspersed in approximately equal numbers with those having neutral water; however, westward of Lake Tyrrell the bulk of the bores have acidic waters. During a 50-hour pumping test carried out on the Parilla Sand aquifer at the Bimbourie 4/5 piezometer nest situated about 1 km west of Lake Tyrrell (Figure 7.5a), pH values measured at the pump head remained virtually constant at about 3.6 to 3.7 throughout. Pumping rates were 1.81/sec. Within the Tyrrell Basin the salinity of the acid groundwater ranges from 30,000 mg/l to 270,000 mg/l, although the more highly saline groundwater brines in excess of 200,000 mg/l are generally neutral.

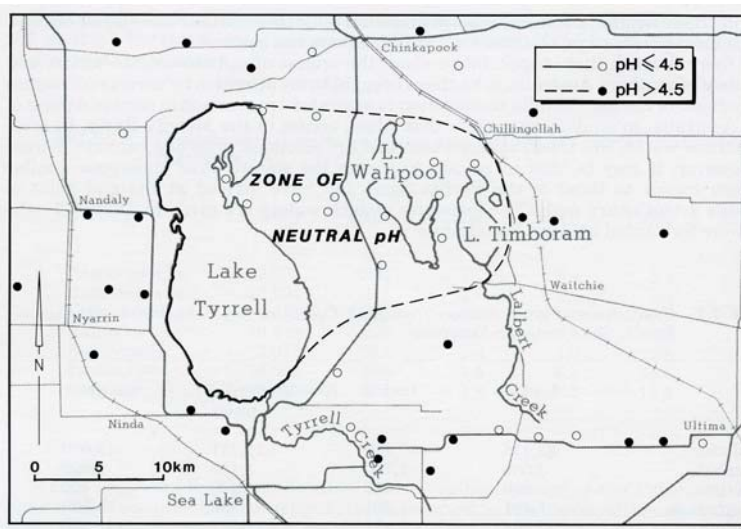


Figure 7.3 - The pH distribution in the groundwaters of the Tyrrell Basin and in the spring waters emerging along the shoreline of Lake Tyrrell and the Tyrrell Creek.

Note that while acid groundwaters emerge along the western shoreline of the lake, neutral or slightly alkaline water occurs in the springs of the eastern shoreline.

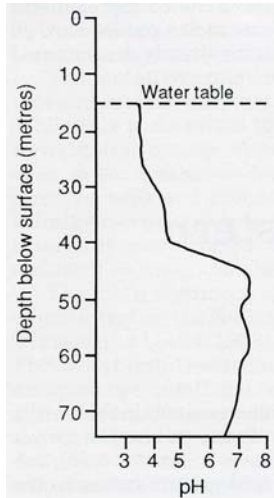


Figure 7.4 - Depth versus pH, Bimbourie 3 bore.

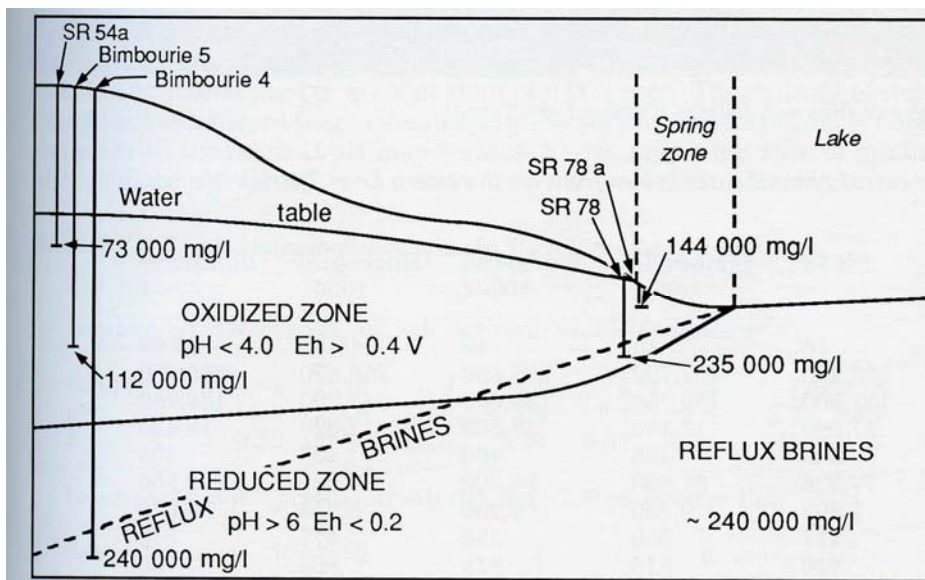


Figure 7.5 - The distribution of pH and redox zonation under the western shoreline of Lake Tyrrell in the vicinity of the T-P piezometer line.

The section shows a general pattern whereby the upper waters in the aquifer have low pH values and are relatively oxidized, while the deeper waters have neutral pH values and lower Eh values. The acid groundwaters do not persist beneath the lake, and, with the exception of a narrow belt seen in the SR 78 piezometer, the reflux brines are neutral.



Figure 7.6 - Tyrrell Creek salt flats and delta as it enters Lake Tyrrell.

TABLE 7.8 Acid and neutral groundwater brines from northwestern Lake Tyrrell

Bore	SR 78	Bimbourie 10002	Bimbourie 10002	Bimbourie 10001	Bimbourie 4
Depth (m)	16	10	10	45-55	65-69
TDS (mg/l)	235,280	267,790	265,350	256,670	245,260
Chloride	131,300	150,750	149,000	144,200	136,750
Sulphate	17,090	18,150	18,500	17,820	18,420
Bromide	389	455	460	402	435
Sodium	77,326	87,800	86,800	84,500	79,500
Magnesium	7,892	9,250	9,200	8,358	8,670
Calcium	417	550	550	477	640
Potassium	710	810	815	806	750
Iron (soluble)	39	15	25	2	1
pH	3.4	3.0	3.9	7.4	7.6

The Bimbourie 10002 analyses were taken 17 days apart, while continuously pumping.

The acid groundwaters are restricted to the upper parts of the Parilla Sand aquifer, occurring at depths to about 25 m below the water table. The lower part of the aquifer generally contains neutral water (Figure 7.4). Coinciding with the pH zonation there is a redox zonation, with the zone of acid groundwaters being more oxidized than the lower neutral waters.

Given the low hydraulic conductivity of the aquifer (0.1 to 10 m/d) and the very low hydraulic gradient (approximately 2×10^{-4}), the residence time for groundwater in the aquifer is quite considerable, and the question arises as to whether the acidity was produced under conditions previously existing in the aquifer, or whether it represents hydrological conditions and hydrochemical processes acting at present.

While it may be that the acidity within the aquifer is of some antiquity, there is strong evidence that there is ongoing superimposition of acidity on previously neutral groundwaters. One line of evidence stems from the presence of a narrow belt beneath the western shoreline of Lake Tyrrell, where comparatively recently formed groundwater brines (Tyrrell reflux brines TDS > 200,000 mg/l) are acid. Both upbasin and downbasin from this belt the brines are neutral (Figure 7.5, piezometer SR 78; and Table 7.8). Since the brines have travelled at most only a few hundred metres from their lake source, the processes creating the acidity have acted within the aquifer,

virtually at the point where the brines are now encountered.

Analyses of acid groundwater brines from the northwestern shoreline of Lake Tyrrell are given in Table 7.8 (bores Bimbourie 10002 and SR 78). For comparative purposes two further analyses of groundwater brines are given which have neutral pH. The neutral brines occur deeper within the aquifer, in one instance (Bimbourie 10001) from immediately below the acid brine, and in the other from the Bimbourie 4 bore situated about 1 km west of the lake (for location see Figure 7.5a). Note the significantly higher soluble iron content of the acid groundwaters.

7.6.2 Acid Springwaters

Where regional groundwaters discharge into the lakes, the springs (Figs 9.1 and 9.2) commonly have low pH values, as do spring fed flats and lagoons. For instance, springs feeding Lake Tyrrell from the northwest, west and south are almost universally acid, with pH values ranging from about 4.5 down to 2.6 (Table 7.9). The salinity of these outflowing waters ranges from about 40,000 mg/l to greater than 300,000 mg/l. The salinity of the springwaters shows a small but significant increase caused by evaporitic concentration. Where the acid spring zone merges with the neutral pH lake waters, a varying wide zone of mixing occurs, marked by intermediate pH values.

TABLE 7.9 - E_h pH characteristics of Lake Tyrrell acidic springs*

Location (a): SW Tyrrell-on salt flat near SR 85 piezometer										
Sample	1	2	3	4	5	6	7	8	9	10
pH	3.5	3.8	3.8	3.5	3.3	3.8	3.7	4.7	3.8	4.1
E _h	0.52	0.54	0.55	0.58	0.57	0.44	0.45	0.37	0.47	0.32
Location (b): NW Tyrrell-on salt flat near T-P piezometer line										
Sample	1	2	3	4	5	6				
pH	3.2	3.3	3.6	3.1	3.2	4.4				
E _h	0.62	0.60	0.55	0.58	0.54	0.46				
Location (c): Far NW Tyrrell-on salt flat near Blanchetown Clay outcrop										
Sample	1	2	3	4	5					
pH	3.2	3.5	3.5	3.7	3.8					
E _h	0.63	0.59	0.62	0.61	0.58					
Location (d): Tyrrell Creek salt flats near Sea Lake rubbish tip										
Sample	1	2	3	4						
pH	4.1	4.1	4.0	4.0						
E _h	0.49	0.51	0.54	0.46						

* Samples taken from different sites within the spring zone.

In contrast to the above situation, the bulk of the springs on the eastern shore of Lake Tyrrell are fed not by acid regional groundwaters but by neutral waters - the Timboram–Wahpool reflux brines (see Figure 8.1). They consequently show none of the low pH induced characteristics found elsewhere in the Tyrrell spring zone.

One unusual spring related feature is that of the lower Tyrrell Creek, which flows across a broad spring fed saline flat prior to entering the lake (Figure 7.6). For most of its course the creek is generally dry, receiving infrequent freshwater flows from upbasin catchments about once every decade. In recent years flows occurred during 1956, 1964, 1973-75 and for a short period in 1981. Seasonal flow in the creek is restricted to the region immediately upstream from Lake Tyrrell, where the water is mostly comprised of saline base flow plus surface runoff from the adjacent bare spring fed salt flats (Figure 9.3). These flats may be considered as an extended spring zone. The pH of the springs ranges from about 3.5 to 4.2 - location (d), Table 7.9 - while the Tyrrell Creek has a pH of about 4 when not in flood.

Once the flow ceases, the water in the creek is restricted to small pools which gradually shrink as evaporation proceeds. The evaporitic concentration results in high salinities and is accompanied by a decline in pH. Pools in the creek have been noted as having salinities over 330,000 mg/l and pH of 2.6-perhaps one of the most hostile aquatic environments recorded anywhere in Australia (see Table 7.10). A general fall in pH after exposure to the atmosphere is a characteristic of the acid groundwaters. This is seen as a reflection of a lack of buffering by bicarbonate and carbonate species, while redox processes result in the release of H^+ ions into solution.

TABLE 7.10 - Water analyses from pools in the Tyrrell Creek

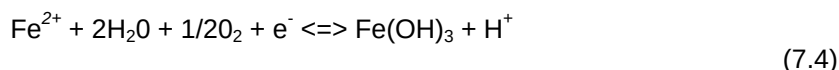
Sample	1	2
Date sampled	4/12/1980	8/2/1980
Position	Chinkapook Road bridge	Ford, downstream from Chinkapook Road bridge
TDS (mg/l)	330,700	327,700
Chloride	182,000	185,000
Carbonate	Nil	Nil
Bicarbonate	Nil	Nil
Sulphate	21,309	23,100
Bromide	655	625
Calcium	61	270
Magnesium	13,590	12,500
Sodium	111,800	105,000
Iron (soluble)	200	140
Silica	54	
pH	2.3	2.75

7.6.3 Increases in Acidity on Storage

One notable feature of the pH measurements is their fall in value during storage in polyethylene bottles: values 2.7 to 3.0 were not uncommon and one value was as low as 2.1. All the above readings were from samples that had remained in storage for up to nine months prior to rechecking the pH. An indication of the extent of pH change with storage time can be had from the analysis of a number of samples from the bore Bourka 8007 (SR 85) situated at southwestern Lake Tyrrell. The pH measurements were done on samples stored in the laboratory for periods ranging from 2 to 9 months. The plot shows a general decline in pH from the field value of 4.8 to a value of 3.0 over this period (Figure 7.7). Iron staining is commonly observed on the polyethylene bottles after some time has elapsed.

An investigation of the processes causing changes of pH with storage time was carried out by Rittenhouse et al. (1969), after they noted alterations of pH in waters collected in polyethylene bottles. It was found that the pH of high salinity waters falls and that of low and intermediate salinity waters rises. They concluded that the changes resulted from temperature dependent oxidation reactions, brought about by the ability of the polyethylene bottles to breathe. In explanation they quote H.F. Young, who in a personal communication interpreted the increased acidity as due to oxidation of ferrous to ferric iron, followed by the hydrolysis of the ferric iron.

A more complete explanation was earlier provided by White (1965), who, commenting on the commonly reported range of 4.5 to 6.5 for oil-field brines, noted that special field measurements are required to confirm this data. White notes that a factor that probably accounts for many laboratory determined low pH values involves the access of atmospheric oxygen during and after sampling. In waters having low buffering CO_2 species, such as where high Ca^{2+} content exists, there is often a significant concentration of Fe^{2+} ions. Oxidation of the ferrous iron produces H^+ ions and therefore increased acidity, according to the reaction



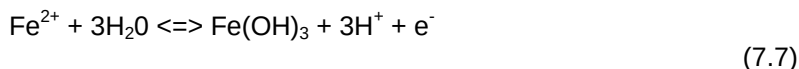
A decline in pH also accompanied the evaporitic concentration of springwaters in small depressions in the marginal flats of Lake Tyrrell. Pools in the bed of Tyrrell Creek had salinities of about 350,000 mg/l, with pH values of 2.6. The fall in pH on evaporitic concentration was investigated in the laboratory, where three acid groundwaters and a sample from a saline pool in the Tyrrell Creek were evaporated in a flask open to the atmosphere (Figure 7.8). All showed a steady fall after evaporation commenced; in sample 1, where evaporation proceeded furthest, the pH fell to about 1.5 - 1.6 and then remained constant despite further evaporation.

Analyses from two pools in the Tyrrell Creek bed are given in Table 7.10. Each sample has an unusually high soluble iron content. The first analysis is that of sample 1 (above), while the second is from a nearby pool. Ironstone, deposited by groundwater, outcrops extensively in the creek bed. Given the high iron content of the groundwater and spring fed Tyrrell Creek water, the explanation given by White (1965) for the decline in pH of water samples would adequately explain the lowering of pH observed in the laboratory and in the field. A more thorough account, whereby the pH of an iron-rich solution falls on aeration, is given by Sato (1960), who suggested that the $\text{H}_2\text{O}_2\text{-O}_2$ couple,



$$E_h \rightleftharpoons 0.682 - 0.591 \text{ pH} \quad (7.6)$$

controls the upper limits of the oxidation potential, where a ferrous solution is subject to oxygenation. This is regardless of the form of the iron present, or whether the solution is homogeneous or heterogeneous. The presence of a solid iron phase developed during aeration causes the pH of the iron solution to be dependent on oxidation potential - the pH falling markedly upon exposure to air. Sato (1960) bubbled air through a ferrous sulphate solution, the pH of which fell until reaching a virtually constant value of about 3.7. Prior to reaching this value the potential followed a line roughly parallel to the potential line of the ferrous-ferric hydroxide couple,



$$E_h \rightleftharpoons 1.057 - 0.177 \text{ pH} - 0.059 \log [\text{Fe}^{2+}] \quad (7.8)$$

On reaching the vicinity of the line formed by the $\text{H}_2\text{O}_2\text{-O}_2$ couple, the pH and E_h stabilized, changing only slightly after a further 20 hours of continuous air bubbling. The sudden decline in the oxidation rate in the vicinity of the $\text{H}_2\text{O}_2\text{-O}_2$ couple was attributed by Sato (1960) to a change in the mechanism of oxidation once the E_h values reached those of the couple. He stresses the role of hydrogen peroxide in oxygen reduction, noting that when the oxidation potential of the ferrous solution exceeds that of the $\text{H}_2\text{O}_2\text{-O}_2$ couple peroxide is catalytically decomposed by the iron, and oxygen becomes no longer available for further oxidation, even though it may be abundant in the solution. Garrels and Christ (1965) comment on the distribution of E_h -pH measurements of natural aqueous environments, as recorded by Baas Becking et al. (1960); they note that the uppermost limits of most measurements of aerated surface waters fall along a line which roughly parallels the upper boundary of water stability. They refer to this line as the 'irreversible oxygen potential' and note that it corresponds to the $\text{H}_2\text{O}_2\text{-O}_2$ stability boundary, previously described by Sato (1960).

The declines in pH observed in evaporating saline pools, with storage time in plastic bottles, and with heating in a flask open to the atmosphere, are examples of an aeration process but with

differing reaction times. If similar pH- E_h relationships exist, and the same controls operate for the Tyrrell waters as they do for the ferrous solutions of Sato, then a lower pH limit can be calculated for the aeration of the waters of any given ferrous concentration.

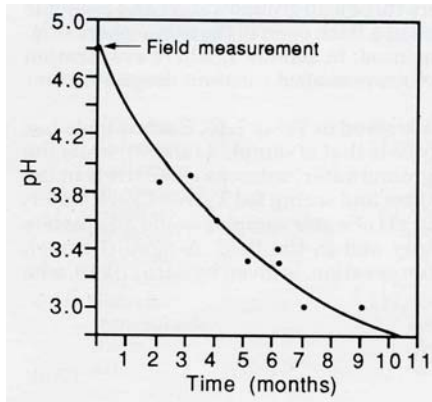


Figure 7.7 - Change in pH, with storage time in poly-ethylene bottles. The samples were all from the Bourka 8007 (SR 85) piezometer at southwestern Lake Tyrrell.

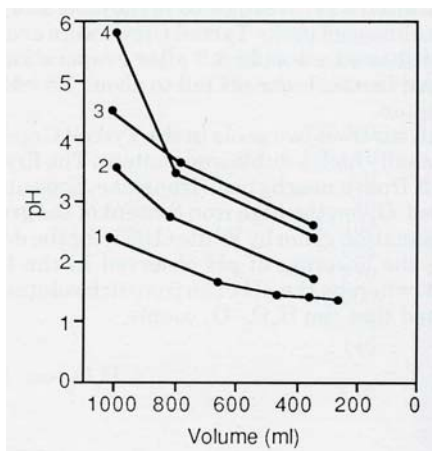


Figure 7.8 - Change in pH of acid waters on laboratory evaporation

1. Pool in Tyrrell Creek—initial salinity 330 g/l
2. Spring (T-P 6), NW Lake Tyrrell—160 g/l
3. Groundwater, Bimbourie 5 bore—112g/l
4. Groundwater, Bimbourie 1 bore—60 g/l

7.6.4 Source of Iron in the Groundwater

Iron is an abundant and widespread constituent of rocks, minerals and soils, being found in sediments, generally in the ferrous form (as the polysulphides pyrite or marcasite and the carbonate siderite), in the ferric form, or the mixed ferrous and ferric form (in various iron oxides and hydroxides, and glauconite). The presence of iron in the groundwaters of the Parilla Sand aquifer may be to some degree attributed to the oxidation of pyrite; however, while this is a source of ferrous iron in the Tyrrell Basin, the most important source of iron stems from the remobilization of iron oxides from the strongly ferruginized Pliocene Karoonda Surface, developed as a weathering cap on the Parilla Sand (Figure 2.27). In the recharge zones on the margins of the Murray Basin and at the inland limits of the marine transgression, the Parilla Sand is commonly heavily ferruginized, the iron oxides and hydroxides being found as dense surface outcrops, as red and yellow coatings on quartz grains, and as thin bands occurring at varying intervals throughout the sequence. The solution of this iron by recharging vadose waters and

through-flowing groundwaters is seen as the major source of iron in the regional Parilla Sand groundwater entering the Tyrrell Basin.

7.6.5 Sources of Acidity in Groundwaters

In natural waters the pH can be influenced by many processes, e.g.

- (1) the dissociation of water



- (2) the reaction of dissolved carbon dioxide in water

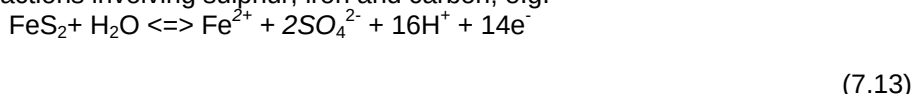


and

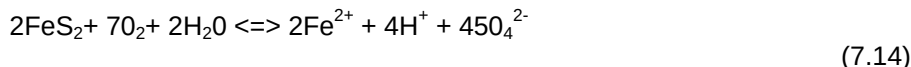


where H^+ is produced in the latter two situations

- (3) oxidation-reduction reactions involving sulphur, iron and carbon; e.g.

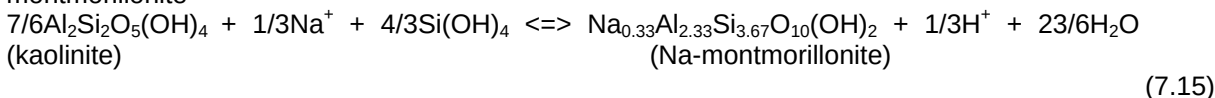


and



- (4) the effects of biological activities such as photosynthesis and respiration, which influence the availability of CO_2 and hence carbonic acid - by (2) above.

- (5) incongruent dissolution of aluminosilicates, for example, the conversion of kaolinite to montmorillonite



This latter process would be most active in the lower parts of the flow path where high sodium and silica are present. While the process may be active in the Parilla Sand aquifer, the tendencies of the groundwaters during their downbasin evolution in the fluvial Loddon Valley sequence was ultimately, on concentration, towards a kaolinite from a montmorillonite (Chapter 5.4.2). Lawrence (1975) notes that the main clay in the upper parts of the Parilla Sand is kaolinite, with some gibbsite present; the kaolinite occupies the interstices between the quartz grains and displays flow textures. Lawrence (1975) also notes that the Parilla Sand waters are in equilibrium with kaolinite. Herczeg (1977), on the other hand, notes that the bulk of 27 samples collected by him have an equilibrium lying on or close to the kaolinite-montmorillonite boundary. However, these samples come from the recharge zone at the edge of the Murray Basin and are relatively fresh, with an average Na^+ content of only 609 mg/l compared with the average of 10,600 mg/l used in the silicate stability diagram (Figure 7.1). Furthermore, in only two instances were pH values of the fresher waters less than 6.0. In general, the acid regional groundwaters from the Tyrrell Basin and surroundings regions tend to be in equilibrium with kaolinite (Figure 7.1), for, although the high Na^+ and SiO_2 contents are conducive to montmorillonite formation (or even in equilibrium with Na-feldspar), the high H^+ (low pH) pushes the equilibrium into the kaolinite field.

(6) Decreased pH with increased ionic strength (seawater). It has been previously observed that, with increased ionic strength of seawater, there is a decrease in pH in response to decreases in the activity coefficients of the CO_3^{2-} ions (Pytkowicz, 1965). This follows from the equation

$$\text{pH} \rightleftharpoons 8.34 + 1/2 \log \text{CO}_3^{2-} \text{ (Krumgalz, 1980)}$$

(CO_3^{2-} is the activity of CO_3^{2-})

(7.16)

Krumgalz (1980) advocates this process as the cause of low pH values (5.9 - 6.7) in the Dead Sea. It may explain similar values in some Victorian salt lakes. However, the extremely low values (down to pH 2.6 in the springwaters and an even lower value of 1.6 reached on evaporation of the acid Tyrrell Creek water) are well below those attainable even with saturated carbonic acid: hence, a mineral acidity is indicated (cf. Stumm and Morgan, 1970). While the above processes may all contribute to the pH of natural waters, in general, the pH of any individual water or group is usually the result of the dominance of a single overriding chemical equilibrium, or a set of equilibria. Of the processes listed above, the most likely to produce an inorganic acid are oxidation-reduction reactions and the incongruent dissolution of aluminosilicates. It is probable that both of these processes are varyingly active in the aquifer. However, of these two, the more likely to produce acid groundwaters in the upper parts of the unconfined aquifer, while leaving the lower parts neutral, is a redox process—this is supported by the concomitant development of redox zonation within the aquifer. In the following account, it is suggested that redox reactions play a significant part in the development of acidity in the groundwater, springwater, lagoon water and creek water of the Tyrrell Basin.

The pH values obtained for the regional Parilla Sand water are significantly lower than those obtained for any other water of the Murray Basin (cf. Lawrence, 1975). The pH range, from 3 to 4.5, is shown by Baas Becking et al. (1960) as occurring in the provinces of thermal waters, high moor peat bogs, acid soils and mine waters. Groundwaters were not shown as having these low values; however, acid groundwaters have been previously described, usually having high iron content and associated with sulphide minerals (e.g. Back and Barnes, 1965). Wherever acid waters are found, they have in common the presence of oxidizable organic or inorganic materials.

The initial low pH values, and further rapid lowering of pH values on boiling indicate that an inorganic acid is present (probably sulphuric acid). The production of similar acid waters in mines has been described by Singer and Stumm, (1970). Acidity arises from the oxidation of sulphide minerals (usually pyrite), which is accompanied by the production of sulphate and the solubilization of metallic compounds. This type of process was inferred by Langmuir (1969b) for the production of low pH values in groundwaters from Maryland. In a study of iron in groundwater, Langmuir compared waters from New Jersey with waters from the Rariton and Magothy Formations of Maryland, which had been previously described by Back and Barnes (1965), and Barnes and Back (1964). The Maryland waters are similar to the Parilla Sand water in having an inordinate number of lower than normal pH values. Langmuir (1969) comments that, the "typically lower E_h and pH values, and higher ferrous iron contents given in Table 4 (the Maryland waters) than in Table 1 (New Jersey waters) reflect relatively higher abundances of pyritic lignite in the Maryland sediments, and the fact that all the Maryland wells tap the aquifer system under water table, or, at most, semi-confined conditions".

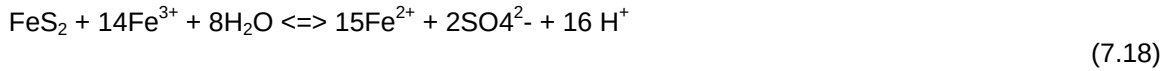
7.6.6 Production of Acidity in Mine Water

The oxidation of pyrite and the release of acidity in mine waters takes place by several steps (Stumm and Morgan, 1970):

(1) The abiotic oxidation of sulphide to sulphate, either by direct reaction with oxygen



or by the oxidation of pyrite by ferric ion



and (2) oxidation of ferrous ion to ferric



The ferric ion formed by this reaction may hydrolyze to give further acidity and insoluble Fe(OH)₃



Comments:

- (1) The rate of oxidation of ferrous ion is significantly slower than that of the oxidation of pyrite by ferric ion, and therefore it is the rate determining step (Singer and Stumm, 1970).
- (2) The oxidation of ferrous to ferric may be mediated by microbiotic influences; for instance, *Thiobacillus ferrooxidans* can catalyze the ferrous–ferric oxidation rate by a factor of 10^5 – 10^6 more than for the corresponding abiotic rate. This is most active in the pH range 2–4.5. However, limitations on the role of individual bacteria species are imposed by the extent to which anaerobic conditions prevail within the ground-water regime. Thus, while *Thiobacillus thiooxidans* is an active oxidizer of sulphide to sulphate in optimum pH conditions (around pH 2), it is a strict aerobe. *T. ferrooxidans* is also an aerobe with optimum pH growth range of 2.5–2.8, but it can utilize the oxidation of ferrous iron as a source of energy (Trudinger, 1967).

7.7 Redox Zonation within the Parilla Sand Aquifer

7.7.1 Oxidation - Reduction Potential

The oxidation-reduction or redox potential (E_h) is a measure of the oxidizing or reducing capacity of a solution. In practice, relatively few elements contribute significantly to the redox processes (apart from oxygen); these include carbon, sulphur, iron, manganese and nitrogen. In groundwater, the measured E_h is often a combination of the various individual contributions. Apart from direct field measurements, the E_h can be calculated from analytical data (Stumm and Morgan, 1970) using the Nernst equation; however, these results can be considerably different from that measured. This arises from the pH and ionic concentrations being out of redox equilibrium with the measured value, often brought about by the presence of organic matter which may significantly lower the E_h while maintaining a low pH.

7.7.2 Redox Potential and pH in the Parilla Sands

Groundwaters in the Tyrrell Basin may be categorized into three distinct groups, based on the redox potential and pH.

- (1) Acid waters, with positive E_h ranging in value up to +0.70V
- (2) Neutral waters, with positive E_h
- (3) Neutral waters, with negative E_h

The vertical distribution of these water groups is seen in the closely drilled areas to the west of Lake Tyrrell, where the deepest bores show neutral waters with low positive E_h values, which pass upward into acid waters with significantly higher positive redox potentials (Figure 7.5). This was especially clear in the piezometer nest drilled at Daytrap Corner, to the northwest of the lake. Here, three bores were drilled to varying depths: Bimbourie 4, depth 64-68 m, Bimbourie 5, depth 23-29 m; and SR 54, depth 16 m. The difference in redox potential between waters at 65 m and at 29 m was 0.5 volt, with a pH range from 7.6 at 65 m to 3.7 at 30 m (Table 7.11). All readings were taken below the water table, within the Parilla Sand aquifer.

TABLE 7.11 E_h pH characteristics of the Bimbourie 4/5 piezometer nest

Bore	Depth (m)	pH	E_h (volt)	Position
SR 54	16	4.4	+0.50	Daytrap Corner, 1 km west of Lake Tyrrell (Figure 7.5a)
Bimbourie 5	23-29	3.7	+0.48	
Bimbourie 4	64-68	7.6	-0.02	

Depth to the water table is 15.2 m.

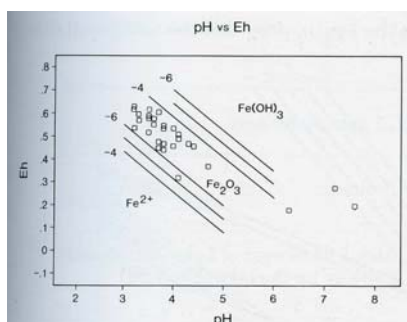


Figure 7.9 – E_h -pH characteristics of springs and groundwaters in the vicinity of Lake Tyrrell. It does not include springs on the eastern shoreline.

A further piezometer nest, situated about 4 km to the south, provided the additional data:

TABLE 7.12 - E_h pH characteristics of the Bimbourie 2/3 piezometer nest

Bore	Depth (m)	pH	E_h (volt)	Position
SR 55B	17	4.3	+0.47	About 1 km west of Lake Tyrrell and halfway up the lake (Figure 7.5a)
Bimbourie 2	51	7.2	+0.28	
Bimbourie 3	72-78	6.3	+0.18	

Depth to the water table is 16.3 m

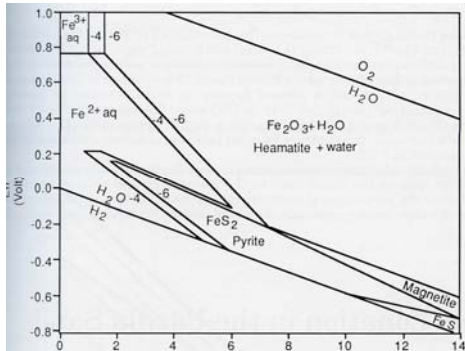


Figure 7.10 - Stability relations of iron oxides and sulphides -
Dissolved sulphur activity is $10^{-2.4}$
Dissolved iron activity is 10^{-6}
(10^{-4} contour is also shown)
At 25°C and 1 atmosphere total pressure

Although only a few readings have been taken to the east of Lake Tyrrell (e.g. the Lianiduck 2 bore at a depth of 50 m, which has a strong H_2S smell and a strong negative E_h), the results give a similar picture, indicating that the same relationship of deeper reducing conditions and shallower oxidizing conditions exists across the Tyrrell Basin. This situation is similar to that described by Sato (1960), who showed a general decrease in redox potential on passing downward from the 'weathering' zone (E_h +0.75 to +0.10V) above the water table, firstly into a 'transitional' zone (E_h +0.5 to -0.15V) and then into a 'depth' environment (E_h +0.1 to - 3.5V)-both below the water table. The tendency for redox potential to decrease with depth was also noted by Germanov et al. (1958).

The presence of positive redox potentials extending to some depth is readily understood in an unconfined aquifer open at the water table to the atmosphere and to local recharge. Edmunds (1973) showed the persistence of oxidizing conditions for 12 km downbasin of the recharge area, after the Lincolnshire Limestone aquifer of eastern England became confined.

7.8 Colour Zonation in the Parilla Sand

Within the sediments of the Tyrrell Basin there is a strong colour zonation shown by a sharp colour change within the Parilla Sand from bright orange or orange - yellow to dull grey or dark grey. These colours are often imparted to the groundwater when piezometers are bailed out. The colour change bears no relationship to the present water table, being generally found at depths of from 2 to 18 m below the water table. To the west of Tyrrell, the boundary is met at a fairly constant depth, but dips down to pass unbroken beneath the lake at a depth of about 12 to 18 m (Figure 2.30). The colour distinction between the zones is essentially due to the presence (or absence) of iron oxide coatings on the sand grains and to the presence of organic matter. The existence of sand which is strongly coated with iron oxide together with acid groundwater in the uppermost parts of the aquifer, closest to the water table indicates stronger oxidizing conditions than deeper in the aquifer, where neutral pH conditions commonly prevail.

Along the southwestern shoreline of Lake Tyrrell the ferruginous zone is about 12 m thick and passes upward into dense spring deposited ironstone at the surface. In a non-steady state environment, where water table and lake level fluctuations have resulted in alternations of recharge and discharge conditions, it is difficult to be certain which events are recorded by the upper ferruginized zone. For instance, it could be assumed that the ferruginous zone is simply a further oxidized zone developed under the present hydrological regime in the uppermost parts of the aquifer closest to the water table, where there is more ready access to atmospheric oxygen; it could also arise from a number of other conditions, such as those occurring during phases of lower water tables; or perhaps during periods of higher lake level, when oxygen-rich recharge water might be expected to enter the upper aquifer. It most probably represents a combination of conditions, its degree of development and preservation being a function of a vertically fluctuating redox and pH zonation in the aquifer, which in turn reflects a dynamic hydrological regime existing through time.

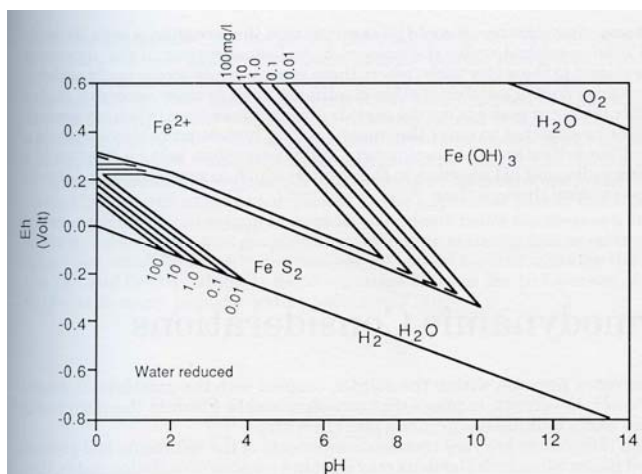
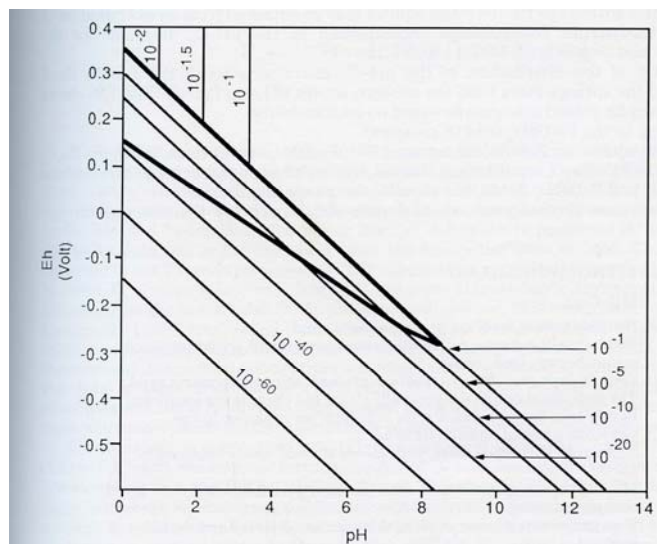


Figure 7.11 - Activity contours of dissolved iron in equilibrium with other solid phase species in the presence of dissolved sulphur.

Figure 7.12 - Concentration contours of SO_4^{2-} in equilibrium with other sulphur containing species in aqueous solution—at 25°C, 1 atmosphere total pressure and total sulphur concentration of 0.1 molar (from Garrels and Naeser, 1958).



7.9 Thermodynamic Considerations

The development of redox zonation within the aquifer, coupled with the considerable degree of vertical groundwater movement, requires that consideration be given to thermodynamic equilibria which are likely to influence groundwater chemistry.

Sulphur and iron compounds are very common components of the sediments and ground-waters of the Tyrrell Basin. Since both elements may exist in a number of oxidation states, they are especially susceptible to changes in redox potential, which may in turn lead to acid waters. The behaviour of both elements under differing redox and pH conditions is well documented (e.g. Garrels and Christ, 1965). An indication of the types of redox reaction likely to occur in differing redox zones within the Parilla Sand aquifer may be obtained from an examination of the various thermodynamic relationships encompassed by the pH- E_h diagrams for the elements sulphur and iron (Figs 7.10, 7.11 and 7.12).

An examination of the distribution of the pH- E_h characteristics of the Parilla Sand groundwaters and the springwaters from the western shores of Lake Tyrrell (Figure 7.9) shows that the waters may be placed into groups based on iron stability:

- (1) Waters lying in the $\text{Fe}(\text{OH})_3$ field (3 samples)
- (2) Waters lying within the Fe^{2+} field in terms of $\text{Fe}^{2+}/\text{Fe}(\text{OH})_3$ equilibrium, but in the Fe_2O_3 field with respect to $\text{Fe}^{2+}/\text{Fe}_2\text{O}_3$ equilibrium. That is, waters whose equilibria lie intermediate between the Fe_2O_3 and $\text{Fe}(\text{OH})_3$ fields, but parallel the phase boundaries.

The ferruginized zone generally extends to depths of 2–12 m below the water table.

BOURKA 1: LITHOLOGICAL LOG

Depth (m)	Lithology
0 - 3.0	Reddish brown, medium-grained quartz sand.
3.0 - 12.5	Reddish yellow-brown, medium-grained quartz sand. Individual sand grains heavily iron coated.
12.5 - 28.5	Grey to dark grey, fine- to medium-grained, micaceous quartz sand. The individual grains are generally clean, but there is the occasional lightly iron coated grain. Pyrite is present as irregular lumps.
28.5 - 40.0	Grey fine-grained, micaceous sand.
40.0 - 46.5	Light grey micaceous sand, with pieces of wood partly replaced by pyrite.
46.5 - 48.5	Micaceous, clayey, sandy silt with pieces of pyritized wood.
48.5 - 54.0	Samples missing.
54.0 - 57.5	Fine grained quartzose sand, with minor wood pieces and minor pyrite.
57.5 - 62.5	Dull grey quartzose silt.
62.5 - 67.0	Fine grained micaceous sand.

Redox zonation within the Parilla Sand aquifer of the Tyrrell Basin shows pyrite to be essentially unstable and where present, as in the Bourka 1 bore, it will tend to go into solution as ferrous iron and sulphate with the release of acidity. The extent to which oxidation of ferrous iron occurs is not fully known; however, it has been observed that the uppermost part of the aquifer is often heavily ferruginized (the orange and orange–yellow zonation previously described). The heavily iron oxide coated quartz sand passes downward into dull grey clean uncoated sand, often containing small amounts of pyrite, e.g. the Bourka 1 log previous page.

The groundwaters are generally acid from the water table down to some depth, and it seems

likely that the iron oxide was deposited in the uppermost part of the aquifer in response to either a high dissolved oxygen content for some depth below the present water table or perhaps to lower water tables. Acid groundwaters emerging at the surface as springs have E_h -pH characteristics which lie roughly between the 10^{-1} molar contour marking the boundary between the Fe^{2+} and $Fe(OH)_3$ stability fields (equation 7.7) and the 10^{-6} contour of the Fe^{2+} and Fe_2O_3 fields, as do many phreatic waters (equation 7.20).



$$E_h \rightleftharpoons 0.728 - 0.177 \text{ pH} - 0.059 \log [Fe^{2+}] \quad (7.21)$$

While Whittemore and Langmuir (1975) note that most high iron groundwaters have E_h and pH values lying on the $Fe^{2+}/Fe(OH)_3$ boundary, Back and Barnes (1965), and Barnes and Back (1964), in a study of groundwaters in Maryland (USA) concluded that the thermodynamic equilibrium exists between groundwater and ferric minerals, intermediate between ferric hydroxide and haematite. The upper limit of solubility is predicted by using the $Fe(OH)_3$ solubility data, while for the lower limit the haematite data is used. This is the situation observed in the Tyrrell springwaters. The explanation given by Barnes and Back (1964) is that in lower E_h waters selective leaching of the more soluble ferric oxyhydroxide phases leaves haematite as the last soluble ferric phase to be dissolved. This conclusion was also reached by Langmuir (1969), who added that increased acidity would also lead to solution of the more soluble oxyhydroxides, as demonstrated earlier by Lengweile et al. (1961), who acid leached the more soluble oxyhydroxides from a mixture with goethite. Langmuir (1969b) re-examined the data of Back and Barnes (1965) and showed that, on the basis of E_h and pK, the groundwaters could be divided into distinct groups: those where ferrous iron was in equilibrium with amorphous $Fe(OH)_3$ and those in equilibrium with haematite or goethite.

The ferrous iron content of the springwaters ranges from about 2×10^{-3} to 2×10^{-6} mol/litre (120 to 0.1 mg/l), the average ferrous content of 80 samples taken over a two-year period from one sampling line, the T-P line, being 10 mg/l. This amount of ferrous iron is within the range expected, where ferrous is in equilibrium with oxyhydroxides ranging from haematite/goethite through to solid phase $Fe(OH)_3$, under the pH - E_h conditions existing in the spring zone.

However, not all groundwaters having positive E_h values plot between the Fe_2O_3 and $Fe(OH)_3$ fields and the three waters lying well within the field of $Fe(OH)_3$ (Figure 7.9) are neutral waters, which are in equilibrium with $Fe(OH)_3$.

7.10 The Influence of Non-Steady State Conditions on Groundwater Acidity

In any system with vertical redox zonation ranging across the stability fields of iron and sulphur, significant falls in water tables will impose oxidizing conditions where reducing conditions previously existed. Under these circumstances, acidity will be produced in the aquifer as pyrite is converted to ferrous ion deeper in the aquifer and as ferrous ion is converted to iron oxyhydroxides closer to the water table. If long-term steady state conditions persist, the unstable phases will all eventually convert, and the production of hydrogen ions and acid groundwaters will cease. In southeastern Australia, however, while steady state conditions may have prevailed for lengthy periods, the position of the water table has nevertheless fluctuated dramatically, especially in late Pleistocene times, and is doing so at present. Indeed, rises in the potentiometric surfaces of regional aquifers, at rates of 0.1 m/y to 0.3 m/y, are now being recorded across much of northern Victoria in response to land use changes brought about since European occupation (see Chapter 4). Major hydrological changes occurring over the last 100,000 years are well documented (Chapters 3 and 4) and these changes are reflected at various sites in northwestern Victoria, especially at Lake Tyrrell (Chapter 3.3).

Prior to 32,000 years ago, much of the Tyrrell Basin was under a large palaeolake system, formed by Lake Chillingollah; an expanded ancestral Lake Tyrrell held about 14 m of water (cf. the present day, 0.5 m). Many of the piezometer sites were under water: the Bourka 1 site for instance being under 12 m of water, while the water table adjacent to the lakes was at an equivalently high level. Further away from the lakes, regional water tables were also significantly higher than at present. Since the depth of the redox zones is to some degree determined by their depth below the water table, it is conjectured that water table fluctuations of the magnitude experienced in the Mallee during the last 40,000 years would have resulted in concomitant fluctuations in redox zonation. Thus, a 14 m higher water table would lead to a similar vertical rise in redox zonation, and perhaps favour the precipitation of pyrite deeper within the aquifer as indicated by equations 7.13 and 7.14. Subsequent lowering of the water table would have exposed any pyrite present to oxidizing conditions, with the consequent production of acidity in the aquifer. At the same time, ferric oxides and oxyhydroxides would have been precipitated in the newly exposed vadose zone and in the shallow phreatic zone close to the water table, again with the production of acidity as shown by equations 7.7, 7.19 and 7.20.

To summarize, vertical redox zonation occurs within the unconfined regional aquifer systems of the Murray Basin of southeastern Australia. Climatically controlled cycles of water table fluctuations, acting upon iron and sulphur species occurring in the groundwaters of unconfined regional aquifers (both bedrock and unconsolidated systems), enable the production of hydrogen ions and hence acid groundwaters within these aquifers.

Significant fluctuations in the climate and water budgets of southern Australia which are known to have occurred throughout the Pleistocene period continued well into the Holocene (Macumber, 1978, 1980). Present-day water tables are rising rapidly throughout much of northern Victoria, and the data from Lake Tyrrell given in this study suggest that a fall in lake levels in this region occurred as recently as during the last millennium. Therefore, steady state conditions in the groundwater systems are more short-term illusion rather than longer term reality; and while this situation exists, there is an ongoing capacity for the cyclic production of acidity in the aquifers by redox reactions.

7.11 Iron Deposition in the Spring Zone

Acid groundwaters have the capacity to transport a large amount of ferrous iron, and amounts as high as 150 mg/l of soluble iron have been recorded from bores in the Tyrrell district. More often, however, the amount of soluble iron is less than 10 mg/l. The bulk of the groundwater flowing into a lake enters through a narrow belt along the lake perimeter (McBride and Pfannkuch, 1975), and at Lake Tyrrell there is widespread deposition of iron oxides along the lake shoreline. This arises from two processes: firstly, deposition in the spring zone, where the ferrous rich springwaters are exposed to the atmosphere; and secondly, at the lake edge, where there is a rapid rise in pH as acid springwaters merge with the neutral lake water. The position of the latter boundary varies almost daily in winter in response to fluctuations in the wind regime, thus giving a broad zone along the lake edge over which iron may be deposited in response to a rapid rise in pH.

Ironstone is found extensively in the spring zone and marginal flats of the western and southern shorelines of Lake Tyrrell. The association of iron with acid groundwaters is readily illustrated by the absence of similar iron formations on the eastern shoreline, where the springs are fed by the neutral pH Timbaram - Wahpool reflux brines. In the south and southwest of the lake, where the Parilla Sand outcrops along the shoreline, and on the Tyrrell flats, ironstone is found almost continuously along the lake edge, and at times extends lakewards across the marginal flats for several hundred metres (Figure 7.13). Even where no ironstone is visible at the surface, the underlying sediments are often heavily ferruginized and dense ironstone is generally encountered within a few centimetres of the surface, as in Figure 7.14.

Within the spring zone, concentrations of red iron oxide oozes are frequently associated with outseepage points, while small flat ironstone 'biscuits' are scattered across the surface. In some instances, ironstone polygons have developed, ranging in size between about two and four metres (Figure 7.15). The polygons are associated with desiccation or shrinkage cracks, which provide preferential pathways for upwelling groundwaters. The margins of the polygons are turned up, often forming tepee structures which are caused by the ongoing deposition of iron oxides along the polygon edges. Gypsum and halite are commonly associated with the margins (Figure 7.16), while within the polygons there are a large number of ironstone fragments derived by erosion of the tepees during periods when the lake invades the spring zone.

Where iron is extensively deposited along the Tyrrell shoreline, the resulting ironstone deposits are at times modified by chemical and physical processes (such as wave action), and ironstone shore platforms are developed which may extend across the spring zone and onto the lake floor. These are similar to the shore platforms developed in marine coastal situations. Since the latter areas are also zones of groundwater discharge, it is likely that the process is essentially the same in both coastal and continental settings, with groundwater-deposited iron providing the cementing medium.

In places, ferruginized platforms can be traced stratigraphically beneath lacustrine sediments deposited during the last significant late Pleistocene lake-full phase at Lake Tyrrell. As is the case elsewhere at Lake Tyrrell, the lacustrine deposits are overlain by aeolian gypseous sediments deflated from the lake floor during the last major phase of aeolian activity, which is believed to have occurred ca. 20-16,000 years BP. The deposition of ironstone from inflowing groundwater within the spring zone at earlier times is not unexpected, and merely confirms that iron transport by groundwater was an ongoing process during the Pleistocene. Sections through the shoreline sequences show vertical variations in iron oxide distribution, arising from the expansion and contraction of the spring zone in response to lake level fluctuations over extended periods (Figure 7.17). Under lake-full conditions, the spring zone and the zone of iron oxide deposition are compressed between the rising depression rim and the lake edge; conversely, as the lake shrinks, the spring zone expands, as does the zone of iron deposition which extends across the spring zone and the newly exposed lake floor to the lake edge.



Figure 7.13 - Ironstone deposited by upwelling groundwaters in the spring zone at southwestern Lake Tyrrell.



Figure 7.14 – Ferruginised lacustrine sediments, Raak main salina, Raak boinka



Figure 7.15 - Ironstone polygons in the spring zone of southwestern Lake Tyrrell.



Figure 7.16 - Rim of ironstone polygon, showing coatings of gypsum and halite deposited by the evaporitic concentration of springwaters.



Figure 7.17 - Laminated lake sediments, ferruginized at the top and overlain by aeolian gypseous clays. The ferruginization is caused by groundwater outflow but might be mistaken for a pedogenic feature suggesting subaerial exposure for a significant time between the deposition of the two units.



Figure 7.18 - Contemporary ferruginization of a ground surface. Here the laminated lacustrine sediments are overlain by a currently forming ferruginous horizon. The sediments pass down-ward into a dense ironstone formed in an earlier cycle, and a third older cycle is represented by haematite forming the base of a nearby island - Raak boinka.



Figure 7.19 - This artesian bore, situated on the edge of the main Raak salina, monitors an interval at about 50 m in the Parilla Sand aquifer. Water seeping from a join in the PVC casing has deposited a coating of iron oxide over the casing and on the sides of a drainage ditch. The salina floor is heavily ferruginized, and islands in the distance have an ironstone base, which can be seen in Figure 7.20.

Northwestern Victoria has been an area of extensive lake systems since upper Pliocene times, and it is probable that groundwater discharge accompanied by iron deposition has occurred on the margins of the various lake systems since their inception. Indeed, bright red colours commonly found in the Blanchetown Clay indicate a significant degree of iron mobilization during its deposition. However, at Lake Tyrrell, the incoming of the modern saline groundwater regimes roughly coincides with the Bruhns–Mituyama palaeomagnetic boundary - dating at about 700,000 years - which, in turn, roughly coincides with a switch from the brown and grey clays of the Blanchetown Clay to the laminated gypseous clay of the Tyrrell Beds (Chapter 2.10). On the assumption that the change to saline conditions occurred roughly at the same time throughout northwestern Victoria, it may be assumed that hydrological conditions conducive to the deposition of iron in the spring zones of the regional groundwater discharge basins have existed for at least the last 0.5 million years. This also applies to other groundwater transported minerals, and underlines the significance of the spring zone environment in mineral investigations.

Over much of northwestern Victoria, the regional water tables are periodically close to the surface, as reflected in the very large number of present-day and fossil groundwater discharge features which are scattered throughout the Mallee landscape (Macumber, 1980). Ironstone is commonly associated with many depressions, for example at the Raak boinka, 100 km north-west of Lake Tyrrell, it outcrops on the floors of active salinas, gypsum flats and dry lakes. High iron content is recorded in groundwater which feeds the boinka (Table 10.7), and iron oxide freely precipitates on the ground around flowing bores (Figure 7.19). At the main Raak salina the sandy lake floor sediments are heavily ferruginized, and lake floor remnants, now existing as bedded ironstone pedestals rising about one to two metres above the lake floor, provide a base for aeolian sand islands (Figure 7.20). The remnants of similar pedestals are often found as jumbled heaps of ironstone fragments protruding above the lake floor. Tepee structures are well preserved on the individual thin ironstone beds. In some instances, a further distinct ferruginous horizon commences at a depth of about 30 cm beneath the lake floor, while a third layer is now forming on top of the overlying laminated lake sediments (Figure 7.18).

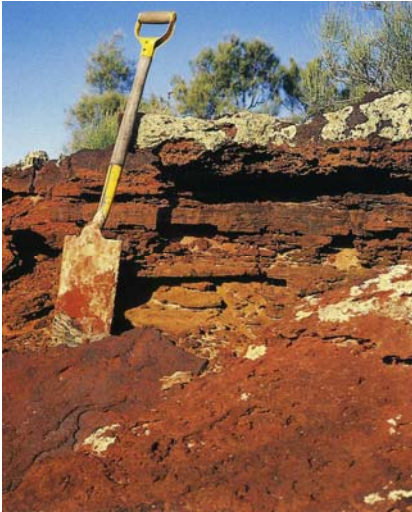


Figure 7.20 - Ferruginized lake sediments forming an ironstone base to aeolian islands in the main Raak salina. The height above the lake floor gives an indication of the extent of lake floor lowering since ferruginization occurred.



Figure 7.21 - Redox zonation developed along the polygonal dessication cracks beneath the halite tepee structures

In the main salina of the Raak boinka, halite tepees form over vertical desiccation cracks after the lake dries out. Removal of the halite tepees shows the cracks to be infilled with red iron oxide, while the surrounding areas are dark coloured, showing reduced conditions; that is, the desiccation cracks are preferential pathways for the movement of iron solutions and evaporite salts towards the surface. The boundary between the red oxidized and dark grey reduced sediments is sharp; it is essentially a redox boundary (Figure 7.21).

At Raak, as at Lake Tyrrell, the development of the ironstone sequence is an ongoing process.

Data from a borehole drilled on the salina edge shows that the ferruginized lake floor sands extend downward for only 3–4 m, at which point a thick clay sequence, (the Blanchetown Clay) is met. Stratigraphically, the iron formation correlates with the post-Blanchetown Clay sequence as at Lake Tyrrell, and is therefore less than 700,000 years old.

In summary, ongoing iron deposition has taken place in the spring zones and on the lake margins of regional groundwater discharge systems of northwestern Victoria for much of the Pleistocene and the Holocene. The main source of the iron in the groundwater is from the re-solution of iron oxides deposited during an earlier Pliocene cycle of ferruginization which saw the development of a lateritic profile—the Karoonda Surface—over much of the western Murray Basin. The widespread distribution of groundwater derived ironstones in depressions throughout the Mallee raises the question to what extent have spring deposited ironstones been mistaken in the past for the Pliocene Karoonda Surface. There is no dispute about the origins of ironstones now found high in the landscape; however, consideration must be given to the likelihood that those deposits found at low points in the landscape, especially in association with playa lakes, represent Pleistocene and Holocene ironstones of groundwater origin.

7.12 A Regional Groundwater Discharge Model for Continental Syngenetic and Epigenetic Ore Deposits

Within zones of regional groundwater discharge, discrete discharge features are commonly found adjacent to large faults lying across the flow path of regional groundwater systems. Examples are the north-south trending Tyrrell Fault, Leaghur Fault and Danyo Fault. These faults disrupt the flow systems, causing the formation of the large discharge basins of northwestern Victoria. The discharge zones and specific discharge points - the lakes and boinkas - have been in existence for a considerable period of time - at least 0.5 million years, and perhaps considerably longer.

Similar disruptions to regional flow occur where coarse-grained continental sedimentary suites are replaced by fine-grained marine sediments in the deltaic areas of the Tertiary Murray Basin (as in the Kerang district), again forcing regional groundwater discharge and the development of a typical groundwater discharge-lacustrine landscape. Regional ground-water outflow is concentrated in the spring zones of the lake systems where, as a consequence, mineral deposition (gypsum, silica, iron, etc.) commonly occurs.

Acid, oxidized groundwaters, such as those contained within the Parilla Sand aquifer, have the capacity to readily transport metal ions such as copper, silver, uranium and nickel from upbasin source areas - whether concentrated or disseminated - into the regional discharge zones. This is clearly the case with ironstone deposition; however, an additional example is the radium which occurs in the spring zone of the T-P line at northwestern Lake Tyrrell. The radium is associated with light coloured aluminosilicates deposited within the spring zone (see Chapter 9.3.7). It occurs where acid groundwaters having pH values of about 3 pass lakewards into a neutral pH zone where extensive reduction of sulphate occurs, forming black sulphide-rich muds. The source is monazite contained in heavy mineral beach and near shore placer deposits which occur scattered throughout the Parilla Sand.

The solution of metal ions is enhanced by the high chlorinity of the regional groundwater, which enables ions such as copper, silver and lead to be transported as chloride complexes. Since they have had long existences, the regional discharge zones in general and the spring zones of salt lakes and boinkas in particular must be regarded as significant sites in continental settings for the concentration of metal ions carried by the groundwater systems.

Deposition of these metals as groundwater epigenetic sulphides and oxides is favoured by marked E_h -pH zonation, both vertically within the aquifer and horizontally across the aquifer as the lake edge is approached. Syngenetic ore bodies could form in the zone of sulphate reduction, commencing at the contact between the spring zone and the lake. It is notable that both epigenetic and syngenetic deposits may form within close proximity to each other; the former being deposited within the aquifer close to the shoreline, while the latter forms at the surface. The deposits may merge vertically, and it follows that in these settings a single ore body may be partially syngenetic and partially epigenetic (as in the case of the ironstones). An especially favourable depositional environment occurs in the stream delta, where the base flow Tyrrell Creek (draining extensive discharge flats, and hence flowing with a pH of 4 and a salinity of 40,000 mg/l) enters the neutral pH lake system. The floor of the creek is covered with ironstone, and the creek dries up in summer, forming saline pools in which halite forms.

In conclusion, the regional groundwater discharge model provides a simple explanation for certain syngenetic/epigenetic ore bodies where an association of red beds, evaporites and base metal sulphides is found.

8. REFLUX BRINES

8.1 Introduction

The Parilla Sand aquifer, beneath the three discharge lakes Tyrrell, Wahpool and Timboram, does not contain the typical 3-5% TDS regional groundwater. Instead over an area covering some 300 square kilometres, the aquifer is filled with brine, varying in salinity from about 9% up to 29% (290,000 mg/l). The brine may be separated into two distinct groups, based upon salinity and location (Figure 8.1).

- (1) Tyrrell brine. The salinity range is from about 200,000—290,000 mg/l. This brine saturates the entire Parilla Sand aquifer beneath Lake Tyrrell, and has been found at depth in bores one kilometre west of the lake, where it underlies the inflowing regional groundwater.
- (2) Timboram—Wahpool brine. The salinity of this system is from about 95,000 mg/l to 125,000 mg/l. It occurs beneath regional groundwater in the area between Lake Wahpool and Lake Tyrrell and directly underlies lakes Timboram and Wahpool.

It is shown in Chapter 12.3.6 that the brines in the Tyrrell Basin have accumulated within the last 32,000 years.

Although having different salinities, the brines and the regional Parilla Sand water are similar in chemical composition, with the major ionic ratios retaining the same strong resemblance to seawater as noted previously in the case of the regional groundwater (Table 8.4). The restriction of the brines to that part of the aquifer beneath the respective lake systems and their downbasin shadow zones (Figure 8.1), when coupled with their compositions, indicates that the brines are derived from lake waters whose chemical composition was largely determined by the chemistry of the inflowing regional groundwater. Following a degree of evaporitic concentration in the playa lake environment, the lake waters infiltrated through the lake floor into the groundwater system to form groundwater brines.

8.2 General

The recharge of saline waters into groundwater systems is a widespread problem in oil producing areas (Fryberger, 1975), where oilfield brines may escape from evaporation pits and contaminate underlying aquifers. However, the downward seepage of naturally occurring hypersaline brines from lakes and lagoons into groundwater systems has also received much attention as a consequence of the process of seepage reflux dolomitization (Adams and Rhodes, 1960). Seepage reflux is the term used by Adams and Rhodes for the downward seepage of hypersaline brines through the floors of coastal lagoons, following the evaporative concentration of inflowing seawater. They used the process to explain dolomitization of underlying limestones by magnesium enriched bitters formed in the lagoons. The process was used by others, notably Berner (1971), Murray (1969), Deffeyes et al. (1965), and Lucia (1972) to explain areas of dolomitization in near coastal situations. In these cases the refluxing brines returned seaward via the groundwater systems, in one instance underflowing less dense seawaters moving landward toward the coastal lake system (Deffeyes et al., 1965).

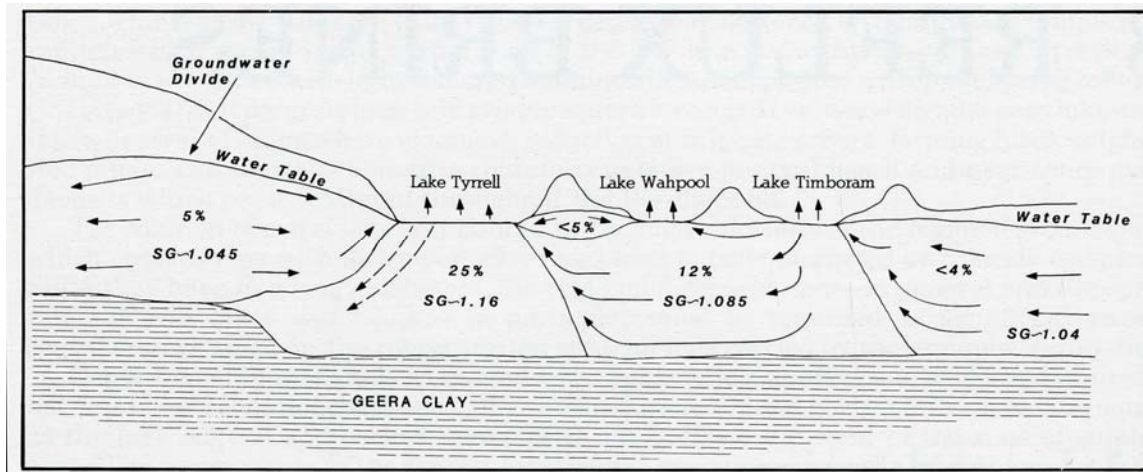


Figure 8.1 - Groundwater flow systems and salinities in the Tyrrell Basin



Figure 8.2 - The Bimbourie 4/5 piezometer nest, looking towards Lake Tyrrell situated about 1 km to the east. At this site, after passing through a broad transition zone, the Tyrrell reflux brines were encountered at a depth of 60 m

Although until now restricted to a coastal setting, the process of brine recharge or reflux is essentially the same for certain southeastern Australian playa lakes and the larger boinkas in their continental setting. Instead of seawater, the water sources are direct precipitation, surface runoff, and groundwater inflow, while the salt source is essentially from the inflowing groundwater. The salt is returned to the groundwater system as a concentrated brine solution after a period of evaporitic concentration in the lake environment. For the purpose of this work, the term reflux or seepage reflux is used to describe the return (or reflux) of brines into an aquifer system, following the evaporitic concentration of salts in a lake which is itself partially fed by groundwater discharge. Brines formed in this manner are termed reflux brines; those beneath Lake Tyrrell are the Tyrrell reflux brines; and those beneath Lake Timboram and Lake Wahpool are called the Timboram–Wahpool reflux brines.

After their return to the aquifer, the brines may again continue to pass downbasin, being gradually diluted by the regional groundwaters as they flow either toward a terminal sump, or, as in the case of this study, toward the Murray River, where they may finally discharge to the surface and eventually be lost to the sea. The downbasin migration of the brines may be interrupted by further cycles of evaporitic concentration in lake systems: this is the case with the Timboram–Wahpool reflux brines, which re-emerge as springs along the eastern shoreline of Lake Tyrrell.

8.3 The Association of Groundwater Brines with Discharge Landscapes

The presence of high salinity groundwaters beneath discharge lakes and complexes is not restricted to the Tyrrell Basin, but is found in similar landscapes throughout northwestern Victoria. Examples of groundwater brines from the Raak and Pink Lakes boinkas are given in Table 8.1. A further analysis, that of the average Timboram–Wahpool reflux brine from the Tyrrell Basin, is added for comparison.

TABLE 8.1 - Groundwater brines from beneath the Raak and Pink Lakes boinkas (mg/l)

	TDS	Cl ⁻	Br ⁻	HCO ₃ ⁻	SO ₄ ²⁻	Ca ²⁺	Mg ₂ ⁺	Na ⁺	K ⁺	Fe ²⁺	SiO ₂	pH
Raak Boinka												
Goonegal 10001	96353	53000	-	221	8230	641	3338	30560	351	1	10	7.3
Bitterang 8001	122170	68000	170	37	8987	550	3800	40250	360	2	10	6.2
Bitterang 8001	123555	68400	-	36	9740	497	3923	40340	408	1	9	6.5
Pink Lakes Boinka												
Mamengarook 1	110000	53750	126	162	7570	325	2510	33000	293	1	3	7.2
Mamengarook 3	106000	52275	127	9	6120	280	2130	31900	320	1	1	5.8
Timboram Reflux Brines												
Average	114312	64642	176	111	8023	815	4735	35460	333	1	13	6.9

8.4 Hydrochemistry of the Tyrrell Reflux Brine

Table 8.2 shows the general chemical character of the Tyrrell reflux brines. It is based on twelve analyses of groundwater, all having salinities greater than 200,000 mg/l. The samples came from varying depths (6 m to 70 m), from bores distributed along the western shoreline of the lake, and from one bore on the eastern edge of the lake. The table clearly shows the close chemical relationship between the reflux brines and the regional groundwater found in the Parilla Sand. It thus adds a further dimension to the conclusion drawn from lake water chemistry see Chapter 101 that virtually all waters in the basin have a chemistry based on that of the regional groundwater.

TABLE 8.2 - Reflux brines from beneath Lake Tyrrell (average of twelve samples with salinity > 200,000 mg/l)

	Regional Parilla water (mg/l)	Reflux brines (mg/l)	Concentration ratio of Parilla water	Ion excess with respect to bromide
Bromide	64.2	347	5.41	1.00
Chloride	19,757	130,956	6.61	1.23
Sulphate	3,188	15,452	4.85	0.89
Calcium	470	514	1.09	0.20
Magnesium	1,353	7,099	5.24	0.97
Sodium	11,077	77,159	6.96	1.29
Potassium	113	660	5.84	1.08
Silica	56	11	0.20	0.04
Bicarbonate*	249	105	0.42	0.08
TDS	36,263	223,549		

*Average of only four samples

The average salinity of the brine in Table 8.2 is about 224,000 mg/l; however, waters furthest from the transition zone around the brine have a salinity closer to 250,000 mg/l, and at times reach 280,000 mg/l. The density of the brine ranges from about 1.14 to 1.19 g/cm³, averaging about 1.16 g/cm³.

The Cl/Br ratio (377), although slightly higher than in the regional waters (Table 8.4), proves that re-solution of a pre-existing halite body is not the source of the brine. Were the latter process operative, virtually no magnesium and very little bromide would be expected (cf. Holster, 1966). Instead, the Cl/Br and Cl/Mg ratios suggest that the primary source of the brine was evaporitic concentration of an oceanic type water (in this case the inflowing regional Parilla Sand water), with some addition of halite (see Chapter 10). The brines, like the regional waters, are essentially, a sodium–magnesium–chloride–sulphate water. The bulk of the ions show a 5 to 7 times concentration when compared with the regional Parilla waters. The most soluble ions, Br⁻, Mg²⁺ and K⁺, are 5.2 to 5.8 times concentrated.

A comparison of the chemistry of the refluxed brine with that of the parent regional groundwater enables some understanding of the geochemical processes operating within the intervening playa lake environment prior to outseepage. Bromide has again been chosen as an index with which to compare the concentrations of other ions with those of the parent regional groundwater. The choice is supported by the similar degrees of concentration shown by other bittern ions (magnesium and potassium), neither of which is significantly precipitated until after halite. The Mg/Br ratio for the Tyrrell reflux brines is much the same as that for the parent regional groundwater (Table 8.4), showing that in the lake environment there was no significant loss of bromide through concentration in organic-rich muds (Mun and Bazilevich, 1962). The Cl /Br and Na/Br ratios show that there is some addition of NaCl to the reflux brine (see below).

Table 8.3 shows the character of the reflux brines at the Bourka 2/3 (4) piezometer nest, south-western Lake Tyrrell, where the Bourka 4 bore was sampled over a 50-hour period while continuously pumping. The salinity changes during this period were insignificant.

TABLE 8.3 - Chemical analyses from the Bourka 4 bore

Time after pump started:	5 minutes	25 hours
TDS (mg/l)	268,003	271,506
Chloride	147,000	146,800
Bromide	343	436
Carbonate	Nil	Nil
Bicarbonate	118	106
Sulphate	23,300	26,900
Nitrate	25	25
Calcium	550	573
Magnesium	8,840	8,880
Sodium	86,700	86,800
Potassium	1,000	1,000
Iron—total	10	6
Iron—Soluble	5.0	2.5
Silica	31	11
pH	6.4	6.6
S.G. at 20°C	1.17	1.17

*Date Sampled, 24.2.81; aquifer level, 17-23 m; yield, 10.43 Us.

On the basis of the bromide ion concentration, there is a major loss of calcium, bicarbonate and silica, and a minor loss of sulphate. The loss of calcium and bicarbonate is readily explained by the formation of evaporites, calcite and gypsum during the process of evaporitic concentration. The formation of gypsum would also account for the lower sulphate concentration, although the loss is much less than that of calcium: this reflects the excess of SO_4^{2-} over Ca^{2+} in the regional groundwater.

As a consequence of the removal of calcium, the Mg/Ca ratio in the reflux brines has been raised to 14:1 (ionic ratio) or 23:1 (molar ratio). The reflux of magnesium rich hypersaline waters is the mechanism proposed for the dolomitization of calcareous sediments in near coastal environments by Adams and Rhodes (1960), Deffeyes et al. (1965), and others previously cited. The Tyrrell data shows that the same reflux process occurs in a continental setting, and all that may have prevented the formation of dolomites within the aquifer beneath the lake is the presence of a siliceous rather than a calcareous aquifer. It is suggested, therefore, that the process of seepage reflux dolomitization need not be restricted to coastal settings, but may occur in semiarid to arid continental settings beneath large playa lakes.

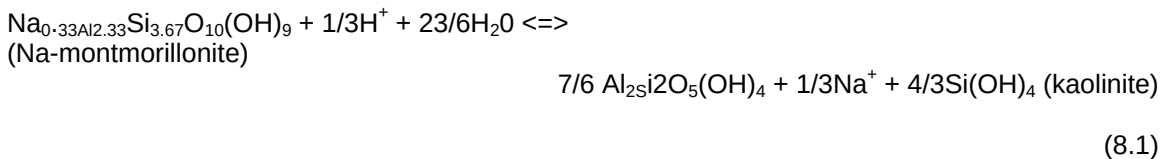
TABLE 8.4 - Chemical ratios—Tyrrell Basin groundwaters

	Cl/Na	Cl/Mg	Cl/Br	Mg/Br	Ca/Br	K/Br	Mg/Ca
Regional groundwater	1.78	14.6	307	21.1	7.3	1.76	2.9
Tyrrell reflux brine	1.70	18.4	377	20.5	1.5	1.90	13.8
Timboram—Wahpool reflux brine	1.82	13.7	367	26.9	4.6	1.89	5.8
Seawater	1.81	14.9	292	20.0	6.2	5.85	2.9

Magnesium—bromide ratios have remained largely unaltered from the ratio in the regional groundwater, while sodium and chloride ions show significant gains. The total NaCl surplus over that in the regional groundwater is about 18.5% of the total Tyrrell reflux brine; this is interpreted as being a non-groundwater component of the brine. One local source of halite is from the lunette complex developed along the entire eastern edge of the lake. It has been previously shown (Macumber, 1968) that halite, deposited with other evaporite minerals in a lunette sequence,

readily undergoes remobilization and collects in low areas adjacent to the lunette.

The decreased Cl/Na ratio, from 1.78 to 1.70, is not considered significant, although this does not preclude the possibility of some minor gain in sodium by interactions with clay minerals, for example (Freeze and Cherry, 1979):



Similar reactions may be responsible for the increase in the potassium content, reflected in the rise of the K/Br ratio from 1.76 to 1.90; at the same time the Mg/K ratio falls from 12.0 to 10.7 (not given in the table). These ratios show that there is no preferential sorption of potassium on lacustrine sediments, either while in the lake or during the reflux process.

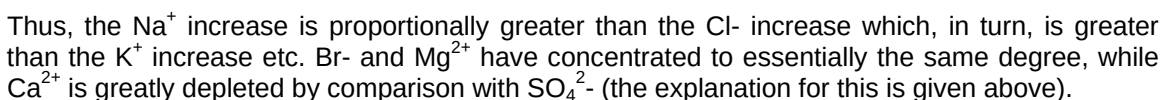
The reflux brine chemistry shows that, while dissolution of clay minerals may provide additional small amounts of Na^+ and K^+ , this addition, in the case of the Tyrrell reflux brines, has very little influence on their gross character. It is concluded therefore that the chemical differences between the Tyrrell reflux brines and the regional groundwater can be explained simply by the normal processes occurring during the evaporative concentration of the regional waters in a lacustrine regime. Some additional sodium chloride has been added, probably from an external source such as the nearby lunette system.

The relatively low silica content of the reflux brine (average 11 mg/l) is in contrast to the relatively high silica content of the regional groundwater, which averages 56 mg/l with values as high as 100 mg/l in some bores. It is normally less than 10 mg/l in the lake water and most commonly less than 5 mg/l. The loss of silica occurs in the spring zone, with the formation of aluminosilicates; this is discussed in detail in Chapter 9.

The Timboram-Wahpool brines are found occupying the Parilla Sand aquifer from Lake Timboram westward beneath Lake Wahpool to the eastern shore of Lake Tyrrell, where they re-emerge as springs. The chemical character of the brine is detailed in Tables 8.4, 8.5 and 8.6. The average salinity of the brine is about 114,000 mg/l, i.e. about three times that of the regional water, but only half that of the Tyrrell reflux brine. Their specific gravity is about 1.085.

Bore	Lianiduck 10001	Lianiduck 10001	Moortworra 10001	Moortwoora 10001	Eureka 10001	Lianiduck 2	Average
Depth (m)		35-48.5	17.5-29	17.5	21-31	48	
TDS (mg/l)	126,900	125,700	120,400	119,700	97,780	95,200	114,300
Chloride	71,005	70,500	69,600	67,400	54,500	53,900	64,600
Bicarbonate	105	244	108	108	18	84	111
Sulphate	8,870	9,140	6,950	8,510	8,020	6,648	8,023
Calcium	1,060	740	865	860	680	687	815
Magnesium	4,825	5,270	4,840	4,835	4,830	3,182	4,735
Sodium	39,600	39,300	37,500	37,400	29,300	29,700	35,500
Potassium	483	310	355	353	245	249	333
Iron-soluble	1	2	1	1	1	1	1
Bromide	185	190	190	189	151	150	176
Silica	17	29	10	10	10	1	13
pH	6.9	7.1	7.1	6.8	6.3	7.1	6.9

Tyrrell brine



In most respects the Timboram-Wahpool brines are similar to the Tyrrell brines-being deficient in calcium, silica and bicarbonate when compared with the regional groundwater but with excess of chloride and sodium (Table 8.6). The same processes that produced the Tyrrell reflux brines are apparently also in operation on the Timboram - Wahpool brines. Notable differences between the two brines are the significant increase in magnesium content, not seen in the Tyrrell brine, and the lesser depletion of calcium.

The smaller calcium loss in the Timboram - Wahpool brines compared to that of the Tyrrell brines can be attributed to lower gypsum losses, reflected in the low sulphate depletion. While gypsum is precipitated in the spring zone and in the lake during seasonal evaporative concentration, the lesser degree of concentration in the case of the Timboram - Wahpool brine, prior to its reflux,

permits more calcium to enter the groundwater system than is the case with the Tyrrell reflux brines. The Mg/Ca ratio of the latter waters is 13.8, compared to 5.8 for the Timboram - Wahpool reflux brine, and 2.9 for the regional groundwater (Table 8.4).

TABLE 8.6 - Timboram—Wahpool reflux brines (average of six samples)

	Regional Parilla water (mg/l)	Timboram— Wahpool (mg/l)	Conc. factor	Excess factor with respect to bromide
Bromide	64.16	176	2.74	1.00
Chloride	19,757	64,642	3.27	1.19
Sulphate	3,188	8,023	2.52	0.96
Calcium	470	815	1.73	0.63
Magnesium	1,353	4,735	3.49	1.27
Sodium	11,077	35,460	3.20	1.17
Potassium	113	333	2.95	1.08
Bicarbonate	249	111	0.45	0.19
Silica	56	13	0.23	0.08

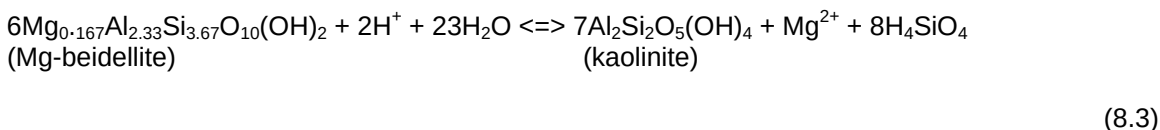
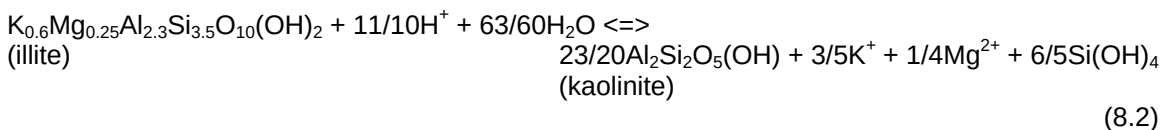
The origin of the excess magnesium can only be speculated upon without further work; however, there is strong evidence that the source is localized and not introduced by the Lalbert Creek, which feeds Lake Timboram from the south. This is indicated by the analyses of Timboram lake water (Table 8.7), which show higher magnesium—bromide ratios (25—27) in the northern part of the lake away from the Lalbert Creek, but normal ratios (19—20) in the south where the creek enters the lake. This difference was also recorded by McLaughlin (1966) in his detailed analyses of saline waters from northwestern Victoria.

TABLE 8.7 - Magnesium—bromide ratios at Lake Timboram

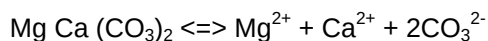
Sampled from	North end	South end
May-Aug. 1980 - (avg of 5 samples)	25.1	20.0
From McLaughlin (1966)	26.7	19.4

While the high Mg/Br ratio may equally well suggest a bromide loss (perhaps by bromide bioconcentrators), the enrichment—depletion figure for the brines shows that the percentage increase in magnesium is the largest of any ion, including sodium and chloride; abnormally low Cl/Mg ratios support this contention. The data indicates an undoubted and significant pickup of magnesium by the infiltrating lake waters prior to, or during, their reflux into the groundwater system. This may be achieved by several processes, for instance the release of magnesium from clay minerals in a manner similar to that for the release of sodium and potassium by incongruent dissolution and by the congruent dissolution of dolomites.

Examples of the first situation are:



The dissolution of dolomite proceeds by:



Silica is produced in the clay reactions; potassium is also released during the conversion of illite to kaolinite, while at the same time hydrogen ion is consumed.

The silica content of Timboram lake water ranges from about 1 to 10 mg/l, averaging about 7 mg/l (Table 10.6); the Timboram reflux brines have a silica content ranging from 1 to 29 mg/l, averaging 13 mg/l (Table 8.6). The higher silica content in the reflux brines may indicate an incongruent dissolution process. It is not known whether or not dolomite occurs within the shallow subsurface sediments beneath Lake Timboram as it does at Lake Tyrrell. Any clue to the possibility of dolomite dissolution, which might have been provided by the increase in calcium, is masked by the overall loss of calcium during the formation of gypsum, both in the spring zone and during evaporative concentration of the lake water. Thus, the Mg/Ca ratio rises from 2.9 in the regional groundwaters to 5.8 in the reflux brines. Therefore, the Timboram–Wahpool reflux brines show a similar evolution to that of the Tyrrell reflux brines. An additional feature, not seen in the Tyrrell brines, is the unusually high magnesium content.

The Timboram–Wahpool reflux brines are chemically very similar to the modern Timboram lake waters, especially during high lake level phases (see Chapter 10, Table 10.6). However, in the case of the Tyrrell reflux brines, the chemistry is quite different to the present-day lake waters (see Chapter 10, Table 10.2). Neither reflux brine shows the re-solution character which is diagnostic of the Lake Tyrrell waters and is reflected in high Cl/Br ratios. Indeed, the bromide versus chloride plot for both the Tyrrell reflux brine and the Timboram–Wahpool brine (Figure 8.3) shows that these waters fall directly on the straight evaporative concentration curve, thus proving that re-solution of halite was not part of their evolutionary process.

8.6 Lake Outseepage and the Formation of Reflux Brines

In terms of chloride ratios, the Tyrrell reflux brines are fundamentally different to the present-day lake waters. This indicates that when the Tyrrell reflux brines seeped from the lake into the aquifer, Lake Tyrrell was not a halite producer as at present (with high Cl/Br), but instead was a groundwater throughflow lake, similar to the present-day Lake Timboram (see below).

While, at times of seasonally high lake levels, vertical hydraulic gradients may occasionally favour outseepage, the low Cl/Br ratios in the shoreline piezometers show that there is no significant present-day outseepage of the halite enriched lake waters (Figure 8.4 a, b). Apart from ephemeral occurrences in very shallow piezometers (< 1.0 m deep), only in the adjacent bores Bourka 8001 and 8005, situated on the lake edge at southwestern Lake Tyrrell, do the Cl/Br ratios occasionally reach values of 500. This compares with ratios of 330 deeper within the aquifer, and over 1000 for lake waters (Figure 8.4 c, d); however, in these two instances the raised Cl/Br ratios are very recent changes and are explained by local recharge of halite enriched runoff water (see Macumber, 1983a).

The chemical evidence indicating that there is no significant lake outseepage is supported by the potentiometric data, since, under the present regime, potentiometric heads within the shoreline piezometers tapping the reflux brine are higher than the level of the lake floor (see Chapter 12). This shows a general inflow of not only the regional groundwaters and waters from within the transition zone but also of the reflux brines (Table 8.8). The lakeward return of the reflux brines is counter to the general outflow over time, and represents a recent shift in the dynamics of the groundwater flow systems; it is manifested by the reversal of the hydraulic gradient within the brines and the closing of the lake to hydrostatically induced outseepage. This is seen to represent only a temporary aberration of the normal outflow condition, while the system readjusts to these recent changes (see Chapter 12).

While Lake Tyrrell is at present virtually closed to hydrostatically induced outseepage, the hydrological equilibrium is delicately balanced, so that, under the large density differentials existing between the reflux brines and the regional groundwaters, only a small fall in closing groundwater pressures, or rise in lake levels, would result in significant outseepage once more.

Most of the readings in Table 8.8 are above the level of the lake floor, which is at a low of 41.3 m along the western shoreline (except off the southwestern shore opposite the Bourka 2/ 3 piezometer nest, where it drops to 41.1 m). The static levels of many bores fall within the range of lake levels obtained during the seasonal cycle of wetting and drying, i.e. from about 41.1 to 41.9 m—the latter level being attained briefly only during the wettest years. However, because of the high densities of the groundwaters, corrections are needed to accurately compare closing potentials and lake floor levels. It is shown in Chapter 12 that average yearly lake levels of 41.5 m to 41.9 m (AHD) are required to overcome the closing groundwater pressures.

TABLE 8.8 - Potentiometric data from bores in the Tyrrell reflux brines*

Bore	Position	S.G.	TDS	Static level (m AHD)
Bimbourie 4	1 km west of lake	1.17	245,000	41.7
Bimbourie 3	1 km west of lake	1.16	240,000	41.9
SR 78	western shoreline	1.16	235,000	41.9
PierMillan 10001	western shoreline	1.16	237,000	42.6
Bourka 8001	western shoreline	1.16	243,000	41.4
Bourka 3	western shoreline	1.17	265,000	41.3
Bourka 2	western shoreline	1.19	290,000	41.1
Bourka 5	western shoreline	1.16	235,000	41.5
Lianiduck 10002	eastern shoreline	1.15	201,000	43.2

* For locality plan, see Figure 7.5a.

It is likely that, in the past, reflux brine accumulation occurred under both slightly wetter conditions (lake high) and drier conditions (lower groundwater divide). On the basis of upward hydraulic gradients now existing within the reflux brines (Tables 12.9, 12.10, 12.11; Chapter 12.12.4), it seems that the most recent hydrological change, leading to the closure of the lake and the development of the thin halite crust, was one which saw a slight fall in average lake levels. For further discussion on these points see Chapter 12.

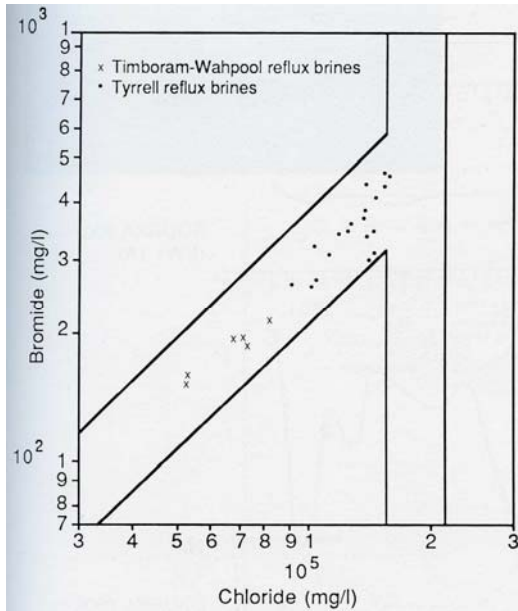


Figure 8.3 - Log bromide versus log chloride for the Tyrrell and Timboram —Wahpool reflux brines.

The distribution of the Tyrrell reflux brines lies in the same field as that of the Timboram—Wahpool brines. This indicates that the reflux brines have evolved from concentrated lake waters which have not undergone a cycle of halite re-solution prior to their outseepage from the lake environment. While this is still largely the case for the Timboram lake waters, it is not so for the present day Tyrrell lake water, which, because of seasonal halite re-solution, always plots in the vertical field of chloride saturation (see also Figures 10.4 and 10.5).

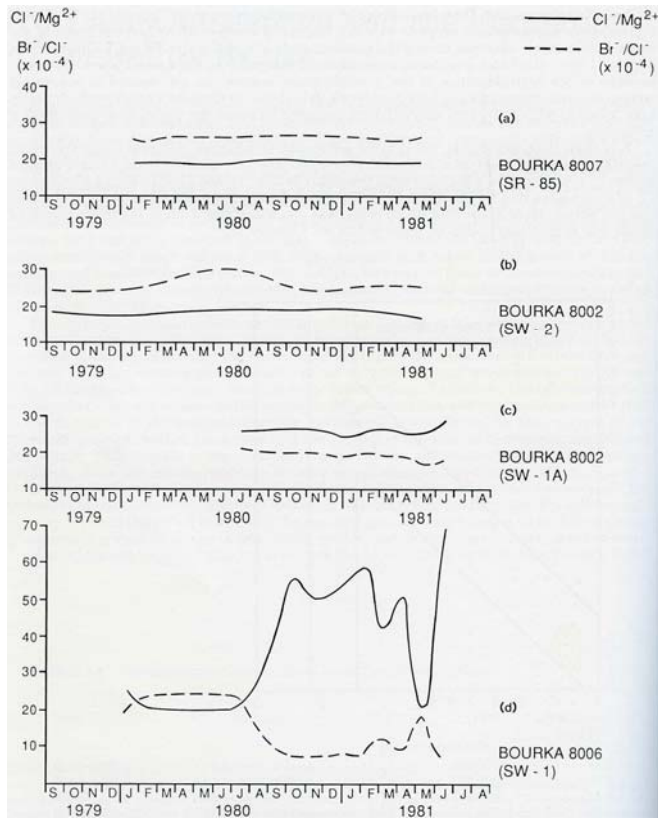


Figure 8.4 - Graphs of $\text{Cl}^-/\text{Mg}^{2+}$ and Br^-/Cl^- in piezometers near the salt works at southwest Lake Tyrrell show variations with time.

The piezometers all monitor the Parilla Sand aquifer at a depth of about 6 m below the water table. Piezometers (b), (c) and (d) are near the lake edge, while (a) is situated about 300 m from the shoreline. Piezometers (a) and (b) show the common (regional) relationships between the various ions, while (c), the closest piezometer to the lake, shows a slightly increased chloride content, which may initially be interpreted as a lacustrine influence. Piezometer (d) shows the regional character up to winter 1980, and then abruptly changes to a Cl^- enriched water. The change is, however, not due to intrusion by lake waters but, instead, to a recharge by halite enriched surface water from a nearby ditch (Figure 8.5).



Figure 8.5 - Bourka 8005 site. The bore with pump attached is in the foreground.

A drainage ditch passes alongside the bore and flanks a halite stack in the background. The water table is about 1 m below ground surface and is readily accessible to the saline drainage waters. This explanation is given for the sudden increased Cl^- content shown in Figure 8.3(d); a similar process may explain the heightened Cl^- content of Bourka 8001, the only other bore to show abnormal ionic ratios, suggesting the possibility of recharge from the nearby lake.

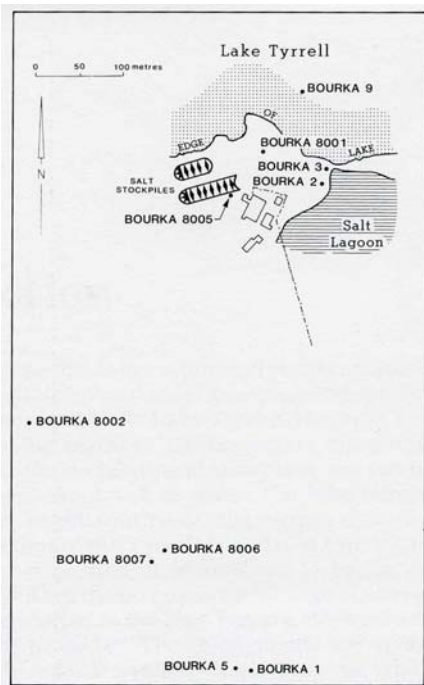


Figure 8.6 - Bore locality plan in the vicinity of the salt works—SW Lake Tyrrell

9. GROUNDWATER INFLOW INTO LAKE TYRRELL THE HYDRO CHEMISTRY OF THE SPRING ZONE

9.1 Introduction

The bulk of groundwater flow into lakes is normally concentrated in a narrow zone around the lake perimeter where a well-developed zone of seepage and spring activity may occur (Figure 9.1) (cf. McBride and Pfannkuch, 1975). At Lake Tyrrell, however, the seepage zone is not always confined to the lake edge. For instance, on its western margin, where the rim of the Tyrrell depression rises sharply to the undulating Mallee plain, the remnants of a low terrace borders the lake. The terrace rises about 1–2 m above the lake floor and is lightly covered with halophytic phreatophytes. In the southwest, the terrace is several hundred metres wide. Its stratigraphy is similar to that of the Tyrrell islands—a thin lacustrine base on which aeolian gypseous clays have been deposited; it is assumed to be the same age as the islands, ca. 16,000–20,000 years BP. Within the terrace are a number of unvegetated salt flat embayments (Figure 9.2) which may be connected to the lake, but are elsewhere cut off by small bars or dunes and hence become lagoons in winter. The embayments are areas of significant groundwater outseepage and as a consequence are undergoing continuous expansion by springhead erosion. Where the terrace is wide and the embayments deep, the bulk of the spring flow is restricted to this zone, with only minor outseepage around the lake edge. For much of the year the Tyrrell Creek salt flats, upstream from its point of entry into the lake, behave essentially as a large embayment, with spring outseepage around the perimeter and groundwater outcrop in the creek bed (Figure 9.3).

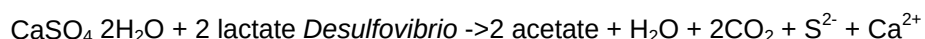
9.2 Groundwater Inflow

The marginal zone of spring activity and outseepage shows a marked seasonal variation in flow. This is partly due to increased groundwater pressures arising from local winter recharge, but is also a result of lower evaporation rates during the cooler months which permit outflow of groundwater that would otherwise be evaporated at the point of outseepage. In many instances spring flow exceeds the evaporation rates even in midsummer; elsewhere this is not the case, and a thin evaporitic crust covers the surface during the summer months (Figure 9.4). In winter, a small rise in pressures within the shallow aquifers, coupled with a decreased evaporation rate, permits a more general groundwater outcrop in the seepage zone which, in turn, leads to shallow water cover over parts of the marginal flats (Figs 9.5 and 9.6).

Two quite distinct groundwaters feed Lake Tyrrell: from the west, south and north come the regional Parilla Sand waters; and from the east flow the more saline Timboram-Wahpool reflux brines. As a consequence, springs fed by the regional waters are generally acid, with pH values often less than 4 (Table 7.9); the eastern springs, however, are neutral and have higher salinities at the springhead.

The inflowing regional groundwaters have a salinity of 35,000–40,000 mg/l. For reasons given below (Chapter 9.3.4) waters approaching this salinity are only found at the uppermost parts of the spring zone—the springhead. On passing across the spring zone, the spring salinity increases in response to an increased salinity in the inflowing groundwaters. The rapid increase in salinity across the spring zone masks the considerable salinity differences existing between the inflowing groundwaters from the west and east—these differences are best seen at or close to the springhead. For example, springhead waters of the eastern shoreline are about 110,000 mg/l, while those on the western shoreline are about 40,000 to 50,000 mg/l (Table 9.2). At the T-P piezometer line (Figure 9.6) on the northwestern shoreline, near-springhead waters at about 60,000–70,000 mg/l reflect higher than normal salinities in the Parilla Sand (analyses SR 54a and T-P 7, Table 9.1).

The chemistry of the outflowing springwaters is reflected in the physical characteristics of the spring zone and marginal flats of the lake. For instance, the surface of the spring zone along the eastern shoreline of Lake Tyrrell is extremely soft: a person rapidly sinks up to or above the ankles (Figure 9.7) in the squelchy, black, iron sulphide rich muds. These muds result from the reduction of gypsum by sulphur reducing bacteria under anoxic conditions (For a fuller account of this process see Teller, Bowler and Macumber, 1982.) Teller et al. note that under saline conditions such as exist in Lake Tyrrell the bacteria *Desulfovibrio* is primarily responsible for the reduction of sulphate (Trudinger, 1976). These bacteria dissociate SO_4^{2-} to acquire oxygen for respiration, and they feed upon organic carbon. This sulphur producing action may be expressed as:



9.1

The black colouration is due to the formation of iron sulphides such as greigite (Fe_3S_4) and mackinawite (FeS). By contrast, the *Desulfovibrio* bacteria cannot survive in the acid spring environments of the western shoreline where positive E_h values also prevail. Thus, the authigenic gypsum growing in the shoreline sediments is not reduced to form iron sulphides and there is consequently little or no black mud layer present. The surface across the western spring zone and the marginal flats is therefore relatively firm by comparison with that of the eastern spring zone. On passing lakeward across the spring zone, a stage of spring zone–lake overlap is reached where, under the seasonal influence of neutral lake waters, pH values become increasingly less acid (Figure 9.8). The rise in pH is echoed by the incoming of conditions conducive to sulphate reduction and the appearance of a black mud layer sometimes accompanied by a narrow zone of algal mat development.

Ironstone is deposited extensively in the spring zone; this aspect has been discussed in Chapter 7.11.



Figure 9.1 - The T-P piezometer line at Lake Tyrrell. Outflowing groundwater with pH of 3.0 passes lakeward, to merge with lake water (pH 6) in the middle ground. This is a period of both high lake levels and high spring flow.



Figure 9.2 - Groundwater outcrop in the spring zone at southwestern Lake Tyrrell. The outflow is from a salt flat embayment formed by springhead erosion of a low terrace bordering the lake.



Figure 9.3 - Groundwater outcrop in the Tyrrell Creek about 200 m upstream from its entrance into Lake Tyrrell. When flowing, the creek at this point normally has a salinity of about 40,000 mg/l and a pH of 4.



Figure 9.4 - The T-P piezometer line across the spring zone at Lake Tyrrell. In summer a thick salt crust accumulates where the spring waters evaporate at their point of outflow.



Figure 9.5 - Groundwater outcrop in the spring zone of southwestern Lake Tyrrell. The flowing piezometer monitors an interval 0.3 m below the ground surface.

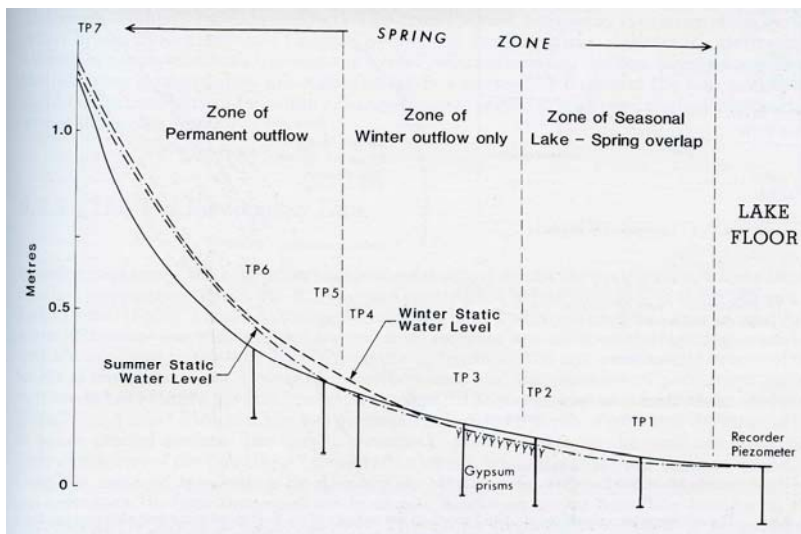


Figure 9.6 - Hydrological zonation across the spring zone at the T-P piezometer site, northwestern Lake Tyrrell. The numbers TP 1-6 refer to the shallow piezometers established to monitor water level and chemical variations with time.



Figure 9.7 - The spring zone on the eastern shoreline of Lake Tyrrell. Springwaters are the neutral pH Timboram-Wahpool reflux brines which are saturated with gypsum, seen here as small raised areas scattered across the spring zone.

Below the surface the gypsum is reduced by *Desulfovibrio* bacteria, resulting in black iron sulphide rich muds, which make the ground surface sloppy to walk across. This is not normally the case on the western shoreline, where gypsum reduction is inhibited by the low pH of the spring water, and the surface is therefore firm.

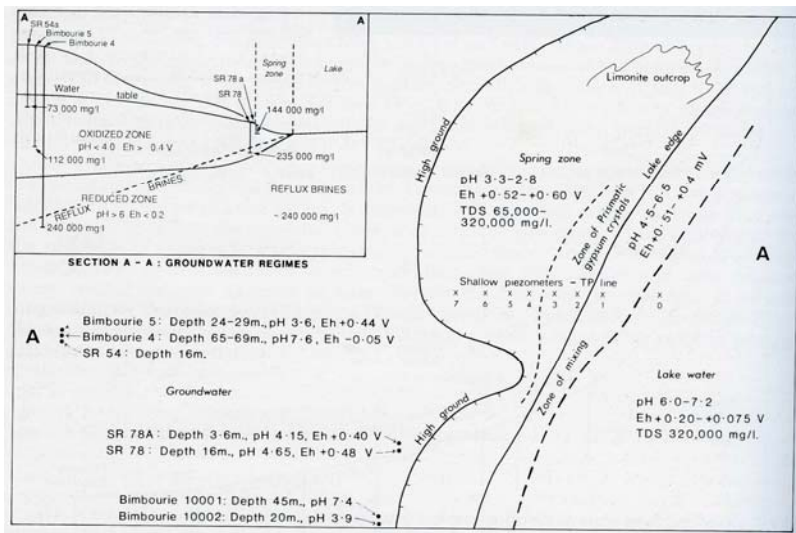


Figure 9.8 - The northwestern shoreline of Lake Tyrrell, in the vicinity of the T-P piezometer line, showing the distribution of piezometers and their pH and Eh values. Note the zone of mixing between the spring and lake waters at the eastern edge of the T-P line. A view along this line, looking shoreward, is given in Figure 9.6. 222

9.3 Hydrochemistry of the Spring Zone

9.3.1 General

The general hydrochemical character of the Lake Tyrrell spring zone was initially determined from a number of samples of outflowing springs at various points around the perimeter of the lake. However, since the spring chemistry shows seasonal and areal variations, it was necessary to sample regularly from a number of shallow piezometers established in lines across the spring zone (e.g. the T-P line on the western shoreline), and out across the lake floor (e.g. the lines SR 6222–6225, and 6226–6229). Greater attention was paid to the western shoreline (in particular the T-P line) because the lower initial salinity of the inflowing regional groundwaters resulted in a wider salinity range within the spring zone; interest was also heightened by the low pH values of the springs.

The chemistry of the spring zone is further complicated by the inflow not only of waters from upbasin, i.e. the regional Parilla Sand water (S.G. I.04) from the west, and the Timboram–Wahpool reflux brines (S.G. I.09) from the east, but also by the return of the Tyrrell reflux brines (S.G. 1.16) - see Chapter 8 - and the various mixes of water of intermediate salinity in a transition zone between the Tyrrell reflux brines and the less dense waters. Since the inflowing groundwaters are concentrated in a narrow strip around the lake perimeter, rapid vertical and horizontal salinity changes were observed in bores and shallow piezometers within the spring zone.

9.3.2 The T-P Piezometer Line

In order to examine the hydrochemical processes acting across the spring zone, a line of eight shallow piezometers (depth 0.5–0.75 m) was constructed across the zone, in the northwest of Lake Tyrrell (Figs 9.1 and 9.8). Seven piezometers (T–P line) were used for periodic sampling, while a Stevens–Leopold continuous water level recorder was set up on the eighth piezometer, which was placed on the lake floor beyond the spring zone. The site was chosen because of the width of the spring zone (several hundred metres) and the presence of permanent spring outflow: it lies within a few hundred metres of the SR 78 piezometer nest on the lake shoreline (Figs 7.5 and 7.5a). This nest has two piezometers—one at a depth of 16 m and the other at 3.6 m below ground surface. The deeper piezometer provides data on the pressure levels and hydrochemistry of the upwelling Tyrrell reflux brines, while the shallower bore provides data from the **zone of transition or dispersion** between the reflux brines and the regional groundwaters. In both instances the hydraulic heads are above lake floor level, with the highest head occurring in the deeper bore. This shows an upward movement of groundwater which appears at the surface in the spring zone adjacent to the piezometer nest (Table 12.9). A further piezometer nest was established at Daytrap Corner, 1 km west of the lake, opposite the T–P spring line. The nest consists of three piezometers, the deepest, Bimbourie 4 (64–68 m), intersected the undiluted reflux brine at about 60 m. Two additional piezometers, Bimbourie 5 and SR 54, provide hydrochemical and hydrostatic data on waters within a zone of dispersion, above the reflux brine. Additional information on the groundwater system was obtained from production bores at the Cheetham Salt Works Site 3 situated on the lake shoreline about 1 km to the south of the T–P spring line (for position of piezometers see Figure 9.8).

9.3.3 Hydrochemistry across the Transition Zone

The transition zone may be considered as a zone of mixed salinities, with the Tyrrell reflux brines at one extreme, and the regional groundwaters at the other. The reflux brines have a salinity of 250,000 mg/l and the regional waters are less than 50,000 mg/l. However, compared to the regional waters the brines are significantly depleted in silica and calcium, which were lost during their previous history of evaporative concentration within the lake environment. As a consequence, the increase in salinity with depth across the transition zone is paralleled by a gradual decrease in total silica: at the same time the total calcium remains virtually constant, there being essentially the same amount in the regional groundwater as there is in the Tyrrell

reflux brine. On a percentage basis, both silica and calcium decrease substantially as salinity increases. Silica shows by far the greatest loss: on the basis of total salts, the silica content in the reflux brines is only 3.5% of that in the regional waters - giving a 96.5% silica depletion in the groundwater across the transition zone.

9.3.4 The Influence of Vertical Salinity Gradients on the Spring Zone

On approaching the shoreline, the flow lines converge as the water table slopes down toward the lake floor, while the saltwater–freshwater interface rises toward the surface (Figure 9.9). The concentration of the inflowing regional groundwater and the transition zone into a decreasing cross-section of the aquifer is reflected in a rapid vertical change in salinity above the interface. On crossing the shoreline, the uppermost parts of the aquifer discharge the least saline water at the springhead, while the deeper, more saline waters continue lakeward to emerge further out across the spring zone (Figure 9.10).

The pattern of vertical salinity distribution within the aquifer is thus passed on to the spring zone, and is readily observed wherever the spring zone is sufficiently wide to allow its detection. An example is given in Table 9.1, where outflowing groundwaters, sampled from the spring zone of northwestern Lake Tyrrell (T-P 7, 5 and 3), are compared with three lakeward flowing groundwaters occurring in vertical sequence at the Bimbourie 4/5 piezometer nest situated about 1 km west of the lake, opposite the spring zone (for locality see Figure 9.8).

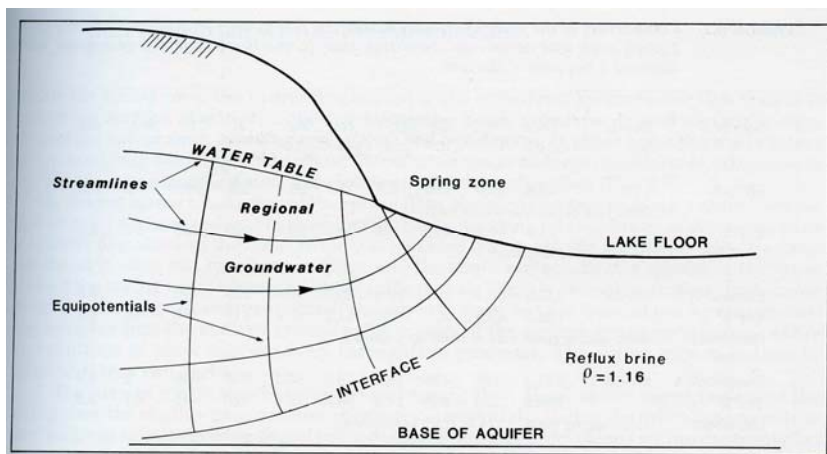


Figure 9.9 - A simplified theoretical flow pattern towards a lake where an interface exists between inflowing regional groundwaters and a reflux brine. Note that the vertical flow section, away from the lake, outcrops within the spring zone in a near horizontal plane.

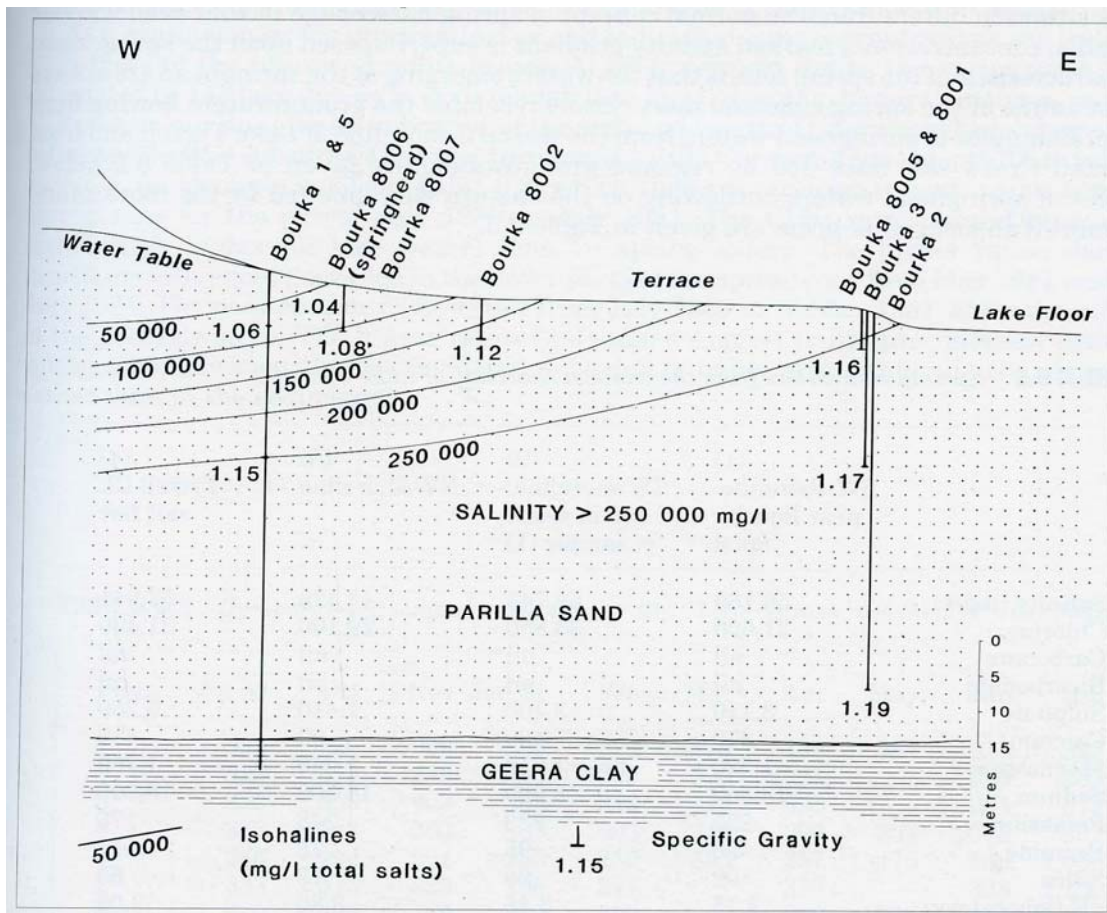


Figure 9.10 – Salinity variation in groundwater beneath the southwestern shoreline of Lake Tyrrell.

TABLE 9.1- Comparison of the hydrochemistry across the T-P spring zone of northwestern Lake Tyrrell with that across the transition zone in the Bimbourie 4-6 piezometer nest, situated 1 km west of the lake

Bore	Depth (m)	TDS $\times 10^{-3}$	Cl ⁻	Br $\times 10^{-3}$	SO ₄ ²⁺ $\times 10^{-3}$	Na ⁺ $\times 10^{-3}$	Mg ²⁺ $\times 10^{-2}$	K ⁺	Ca ²⁺	SiO ₂	pH
SR 54A	17.5	73.6	41.0	120	5.7	23.7	25.0	233	291	71	3.5
	from just below the water table	71.3	40.2	130	4.2	24.2	20.1	290	240	97	3.5
T-P 7 (28/2/80)	from near springhead										
Bimbourie 5	23-29	112.6	60.8	190	10.8	36.0	40.6	320	390	-	3.5
T-P 5 (16/10/80)		126.2	69.3	230	10.5	41.2	42.0	410	275	68	2.8
	from spring zone -91 m from springhead										
Bimbourie 4	64-68	245.3	136	435	18.4	79.5	86.7	750	640	2	6.9
T-P 3 (16/10/80)		238.9	133	458	17.4	78.6	75.6	720	542	67	2.9
	from spring zone -114 m from springhead										

This situation differs from the normal concept of spring outseepage in that even without evaporative concentration a marked salinity gradient is superimposed upon the spring zone.

A characteristic of the spring zone is that the waters emerging at the springhead are always the least saline of the spring zone and most closely resemble the groundwaters flowing from

upbasin. Examples of springhead waters from the western shoreline of Lake Tyrrell and from the Tyrrell Creek salt flats (fed by regional groundwater) are given in Table 9.2 below. Examples of springhead waters outflowing on the eastern shoreline fed by the more saline Timboram-Wahpool reflux brine are given in Table 9.5.

TABLE 9.2 - Springhead waters from the western shoreline of Lake Tyrrell

	(1) SW shoreline, near Bourka 8006	(2) SW shoreline, 200 m south of sample (1)	(3) NW shoreline	(4) Tyrrell Ck salt flat
Salinity (mg/l)	56,169	59,743	48,476	57,941
Chloride	31,600	33,850	29,100	31,400
Carbonate	nil	nil	nil	nil
Bicarbonate	nil	nil	nil	nil
Sulphate	3,770	3,400	1,410	5,250
Calcium	370	460	90	410
Magnesium	1,720	1,710	1,040	1,750
Sodium	18,300	18,900	16,500	18,800
Potassium	225	238	248	175
Bromide	90	95	75	115
Silica	92	90	12	65
pH (laboratory)	3.15	3.45	3.80	3.05

9.3.5 Salt Migration within the Spring Zone

Within the spring zone, the hydraulic gradient of the underlying groundwater flow system is toward the surface. However, while potentiometric heads are above ground surface in some parts of the spring zone, allowing spring flow and outseepage, in other areas the water tables are at, or slightly below, ground surface. These latter areas undergo considerable expansion in dry periods, corresponding with a shrinkage in the area of outflow (Figure 9.6).

On flowing across the surface of the spring flats, the springwaters undergo a rapid increase in salinity as a consequence of both evaporitic concentration and re-solution of evaporitic salts previously deposited on the flats. While this salt load may reach the lake, especially in winter months, it is often lost by return seepage into the shallow groundwater system in the lower parts of the spring zone, where the water table may lie slightly below the surface. In summer months outflowing waters rarely reach the dry lake floor, but are instead lost by evaporation and by reflux into the shallow groundwater system of the central-lower spring zone, where they continue to move lakewards by throughflow processes. The spring zone may thus be subdivided into two parts:

1. The area of continuous outseepage and spring flow. Here, in the upper regions of the spring zone, the shallow groundwater chemistry is essentially that of the inflowing groundwater. The gross salinity gradients and individual ionic gradients developed within the transition zone are reflected in these waters.
2. The lower regions of the spring zone, where water tables are intermittently or permanently subsurface. Here the salts are derived not only from inflowing groundwaters but also from higher up in the spring zone by the processes of throughflow and reflux.

As a consequence, the gross salinities and individual ionic concentrations are well above even those of the returning reflux brines. At its lowermost limits, the spring zone may be inundated by the lake during periods of high lake level. Regular chemical analyses across the T-P piezometer line give an indication of the area and length of time over which lake processes influence shallow subsurface chemical processes within the spring zone (depth 0.5 m below the surface). An

example is given (Table 9.3) which shows a variation in ionic ratios across the spring zone for the period July 1980 to July 1981. The Cl/Br ratio is used because it is significantly higher for lake waters than for spring-waters. The higher ratios show that lacustrine influences prevailed in the lower parts of the spring zone from May 1981 until after July 1981. The near oceanic Cl/Br ratios from July 1980 until May 1981, show the influence of the groundwaters. The lowest ratios in January suggest halite precipitation within the spring zone; however, this is considered to be due to halite precipitation in the piezometers rather than in the sediments.

TABLE 9.3 - Temporal and spatial variations in the Cl /Br ratio across the T -P piezometer line

Piezo	Jul 80	Oct	Nov	Jan 81	Feb	Mar	Apr	May	Jun	Jul
7	333	323	303	303	270	323	323	357	313	323
6	323	313	313	303	323	323	323	-	323	303
5	323	303	303	303	313	323	323	-	294	313
4	313	313	303	303	303	313	303	303	313	303
3	286	294	294	250	250	333	357	476	588	1,000
2	333	313	323	196	227	-	313	-	833	1,250
1	270	323	286	213	200	-	286	909	833	1,111

9.3.6 Silica Distribution within the Spring Zone: The Influence of the Transition Zone

An indication of the influence of the underlying transition zone chemistry on that of the spring zone is seen in the distribution of silica across the spring zone. Tables 8.2 and 8.5 show that, by comparison with the inflowing regional groundwater, silica is greatly depleted in the reflux brines, with both the Tyrrell brines and the Timbora brines showing a similar loss. The lower silica content is also characteristic of lake and lagoon waters throughout the Tyrrell Basin (Table 10.2). An analysis of the silica content of the groundwater within the spring zone was carried out to determine whether silica was lost during transit through this zone. Since, at times, halite is precipitated in the piezometers in the lower regions, bromide was used in preference to chloride to establish concentration ratios with which to compare the silica contents within the spring zone. The plot silica/bromide versus bromide (Figure 9.12) suggests a substantial decrease in silica content in the upper parts of the spring zone as the salinity rises from 71,000 mg/l to about 150,000 mg/l (bromide 250 mg/l). After this point silica loss is still indicated by the slope of the curve, but at a much lower rate. However, caution must be exercised when interpreting this data, since a distinction must be made between silica depletion inherent in the incoming groundwater and silica lost within the spring zone. Several comments can be made on this point:

1. The steeper part of the curve covers that area of the spring zone between piezometers T-P 7 and T-P 5. This area is one of almost continuous outseepage, and the hydrochemistry should clearly reflect that of the transition zone within the underlying aquifer.
2. Lakeward, beyond T-P 5, the water tables are intermittently below the ground surface, permitting a buildup of salts in the shallow groundwater system. A theoretical mixing curve for SiO_2/Br versus Br^- across the transition zone has been added to Figure 9.12. This line is based on average silica and bromide analyses from the springhead, and on their values within the reflux brine, at a depth 16 m below the shoreline, in the SR 78 piezometer (Table 9.4).

TABLE 9.4 - Ionic ratios in the SR 78 piezometer

TDS (mg/l)	SiO ₂	Br ⁻	Mg ²⁺	Ca ²⁺	SiO ₂ /Br	Mg/Ca	Sample date
232,495	14	379	7,860	428	0.037	18.4	28/09/79
235,283	19	389	7,982	417	0.049	19.1	31/10/79

The rough similarity between the steeper part of the springline curve and that of the transition zone curve suggests that the apparent rapid silica depletion in this part of the spring zone is largely an artifact of silica variations within the outcropping transition zone water, rather than a major loss within the spring zone. This is not the case for the lower flatter part of the curve, where salinities occur well above those reached by the inflowing groundwater. Here, a uniform, albeit smaller rate of silica loss within the spring zone is indicated. Whether or not a similar rate of silica loss occurs in the upper areas of the spring zone is more problematical, but it may be that the points to the left of the mixing curve reflect such a loss. Nevertheless, the silica data across the spring line supports the theoretical model that chemical diversity within the spring zone is partly imposed by the diverse character of the inflowing groundwater and partly due to processes active within the spring zone.

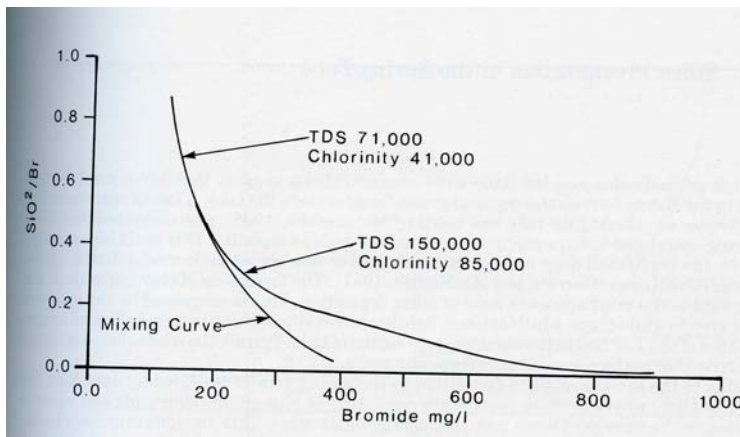


Figure 9.11 – SiO₂/Br⁻ versus Br⁻, T-P piezometer line, Lake Tyrrell

Figure 9.12 - Sharp redox zonation seen in the spring zone in the vicinity of the T-P spring line. This area is highly radioactive due to the presence of radon



9.3.7 Silica Precipitation in the Spring Zone

The high groundwater and low lake silica concentrations suggest that silica precipitation occurs in the spring zone as the regional groundwater enters the lake. A loss of silica from the groundwater on entering the lake was noted by McLaughlin (1966), who suggested that silica was being scavenged to form magnesium rich clays such as sepiolite. This could be accommodated by the very small drop in magnesium; however, the operation is most effective under higher pH conditions (Garrels and MacKenzie, 1967). The formation of clay minerals in the spring zone is the most apparent form of silica deposition. This is suggested by the presence of light grey to white clays, which at times lightly cement silica grains to form ball sized lumps. In the area of the T-P line in the northwestern region of Lake Tyrrell, the white clays are highly radioactive due to the presence of radium and radon.

A clue to the process of silica deposition in the spring zone is provided by data from the eastern Mallee, where a white precipitate consisting of aluminium hydroxide and silica is deposited on the screens of bores pumping acid groundwaters. This precipitate quickly blocks the pump screens. The acid groundwaters are typical Parilla Sand waters, and chemical analysis has shown very high aluminium content; the problem does not occur with nonacid groundwater. Aluminium was not analyzed in the Tyrrell work; however, given the higher solubility of aluminium under low pH conditions, it is accepted that a high aluminium content occurs in the acid groundwaters throughout the Mallee.

It is likely that the white aluminium silicate precipitate occurring in the groundwater pumps is the same or similar to that found in the western spring zone. Precipitation of aluminium (and hence the formation of aluminosilicates) within the spring zone is favoured by the rapid pH rise on crossing the spring zone and during those periods when the lake invades the spring zone.

9.3.8 Hydrochemical Zonation across the Spring Zone

One important consequence of the low pH of the regional groundwater is that despite the rapid salinity increase across the spring zone there is no precipitation of calcite, and instead, calcium remains in solution until it can precipitate as gypsum. This is seen in Figure 9.13, which shows the variations in the Mg/Ca ratio across the spring zone. The ratio rises steeply after salinities of about 290,000 mg/l are reached, coinciding with the development of a zone of prismatic gypsum crystals. The influence of the transition zone on the hydrochemistry of the spring zone is also seen in the case of the Mg/Ca ratio, since, like silica, the Ca/Mg ratio decreases across the transition zone (the Mg/Ca ratio increases). This is reflected in ratios of 18.5 to 19 in the SR 78 piezometer (Table 9.4), while values of about 10 occur at the springhead (Fig 9.13). Therefore, in Figure 9.13 the initial gradient of the Mg/Ca versus TDS graph approximates that of the transition zone mixing curve, and does not indicate calcium loss in the spring zone. The gradient rises sharply only after gypsum commences to precipitate once $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ saturation occurs in the lower regions of the spring zone.

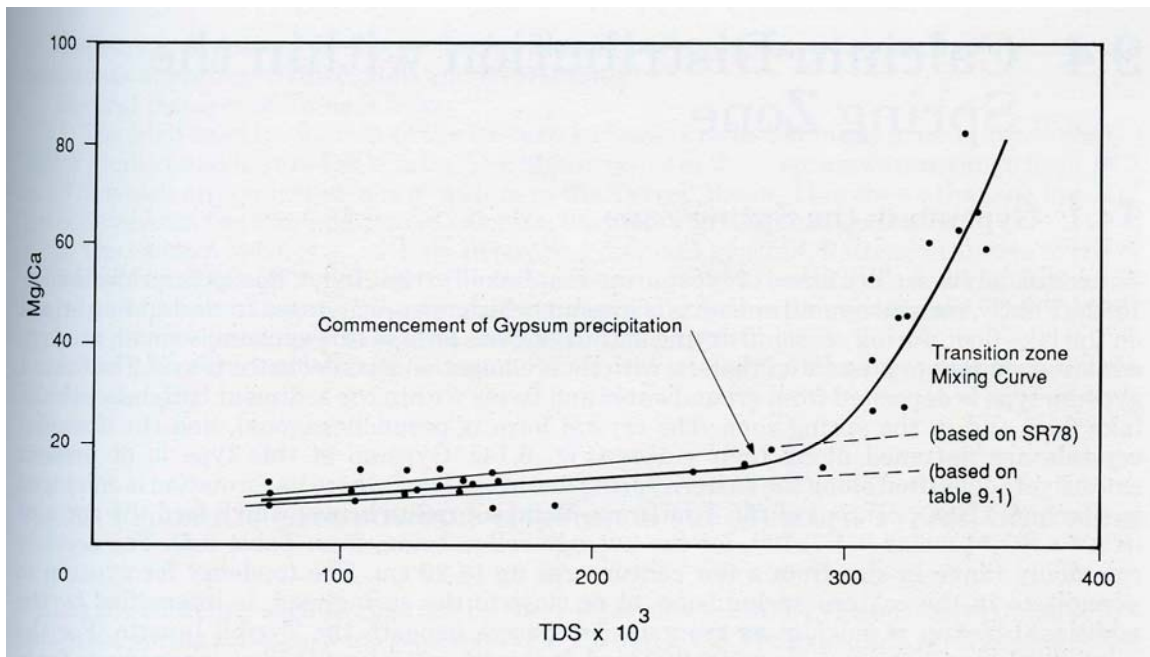


Figure 9.13 - Mg/Ca vs. total dissolved solids, T—P piezometer line NW Lake Tyrrell



Figure 9.14 - Discoidal gypsum precipitated in near surface sediments from inflowing groundwaters within the spring zone (diameter of the coin is 24 mm).

9.4 Calcium Distribution within the Spring Zone

9.4.1 Gypsum in the Spring Zone

Two commonly described forms of gypsum occur at Lake Tyrrell (Teller, Bowler, and Macumber, 1982). Firstly, there are small amounts of gypsum which are precipitated from standing brines on the lake floor during seasonal drying of the lake; the form of this gypsum is small, sand- to silt-sized, equant to prismatic crystals, with their elongation parallel to the c-axis. The second gypsum type is deposited from groundwater and forms within the sediment both beneath the lake floor and in the spring zone. The crystal form is pseudohexagonal, and the discoidal crystals are flattened along their c-axes (Figure 9.14). Gypsum of this type is at present extensively deposited along the eastern spring line of the lake, where its formation is enhanced by the high CaSO_4 content of the Timboram–Wahpool reflux brines which feed the springs, $([\text{Ca}^{2+} \times \text{SO}_4^{2-}])$, molar = 1.7×10^{-3} , for the average reflux brine, from Table 8.6). The crystals commonly range in size from a few centimetres up to 20 cm. The tendency for gypsum to precipitate in the eastern spring zone, at or close to the springhead, is intensified by the additional pickup of calcium as groundwater passes beneath the Tyrrell lunette. For the relatively low salinities of about 100,000 mg/l, it results in $[\text{Ca}^{2+} \times \text{SO}_4^{2-}]$ molar products far in excess of those reached elsewhere at Lake Tyrrell. This is shown in Table 9.5, which gives analyses of four springs from the eastern shoreline, lakeward of the Tyrrell lunette. For comparison, two analyses from the western, shoreline springs are given (T-P 6 and T-P 2). The T-P 6 analysis is of water having a similar total salinity to that of the eastern springs, while T-P 2 is an analysis of springwater from which gypsum is precipitating within the sediments.

TABLE 9.5 - Hydrochemistry of springwaters at Lake Tyrrell

	East Tyrrell springs (194)				West Tyrrell springs (5-6)	
	1 NNE Tyrrell	2 NNE Tyrrell	3 Central Eastern Tyrrell	4 SE Tyrrell	5 T-P 6 NW Tyrrell	6 T-P 2 NW Tyrrell
TDS	103,500	83,200	118,800	117,300	106,750	312,500
Carbonate	41	Nil	Nil	Nil	Nil	Nil
Bicarbonate	46	220	143	229	Nil	Nil
Sulphate	9,340	10,500	10,200	10,800	5,430	20,900
Nitrate	5	5	Nil	5	Nil	9
Calcium	1,615	1,139	1,080	1,250	230	440
Magnesium	2,820	2,208	3,810	4,250	3,520	11,400
Sodium	32,850	26,890	31,300	36,500	35,400	102,100
Potassium	155	95	170	340	390	875
Iron – total	4	2	-	32	-	-
Iron – soluble	2	2	2	2	4	8
Silica	12	10	6	19	83	-
Bromide	155	115	145	215	195	520
pH	8.2	8.0	7.2	7.3	3.0	2.9
S.G.	1.073	1.060	1.086	1.083	1.078	1.210
$[\text{Ca}^{2+} \times \text{SO}_4^{2-}]$ mol	0.0039	0.0031	0.0029	0.0035	0.00043	0.0024
Mg/Ca	1.7	1.9	3.5	3.4	15.3	25.9

Gypsum is deposited on the surface at the point of emergence of spring 4 and within the sediments at springs 3 (discoidal) and 6 (prismatic).

Several features of Table 9.5 are:

1. The high calcium content of the eastern springs is reflected in the greater $[\text{Ca}^{2+} \times \text{SO}_4^{2-}]$ molar product and lower Mg/Ca ratio. The Mg/Br ratios of these springwaters range from 18.2 to 26.3, which are typical values of waters in the Tyrrell Basin. This shows that the

low Mg/ Ca ratio results from an addition of calcium, not from the loss of magnesium.

2. The eastern springs 3 and 4 are depositing discoidal gypsum, flattened relative to the c-axis. However, gypsum deposited at the western spring site 6 consists of prisms up to 6 cm in length, with their elongated c-axes growing vertically within the sediments. This type of gypsum has not previously been described from Lake Tyrrell.
3. The pH of the eastern spring waters is neutral to slightly alkaline, while the western springs are acid.

9.4.2 Prismatic Groundwater Gypsum of the Western Tyrrell Shoreline

Within the shoreline section spanned by the T-P piezometer line, gypsum occurs in the lower part of the spring zone, between piezometers T- P 3 and T-P 2. The gypsum is precipitated below the surface as massed prismatic crystals up to 6 cm long and 1 cm wide by through-flowing groundwaters, which have evolved to the stage of gypsum saturation (analysis 6, Table 9.5). They grow vertically within the sediments (Figure 9.15), being elongate along the c-axis. In many instances the crystals have a cone or wedge shaped hollow core infilled with sand grains. The core is open at its uppermost end, and extends well into the crystal (Figure 9.16).

9.4.3 Physico-Chemical Environment for Prismatic Gypsum Formation

The prismatic gypsum crystals are being precipitated from a concentrated seawater type Na-Mg-Cl-SO₄ solution, with salinity commencing at about 280,000 mg/l, and chlorinity of approximately 150,000 mg/l (twice the chlorinity at which gypsum precipitates from concentrated seawater). Apart from a lower calcium content and higher concentration, the waters are chemically similar to those 120,000 mg/l waters depositing discoidal gypsum on the eastern shoreline (Table 9.5). The waters of the western spring zone are however acid, having a pH range from about 4 to 2.8. The pH in the zone of gypsum precipitation at the T-P site ranges from about 3.2 to 2.8. In winter, when the lake holds water, a rapid change in pH occurs on passing from the spring zone to the lake. Under high lake conditions and an easterly wind, the neutral pH lake waters may occasionally inundate the zone of gypsum development.

9.4.4 Mg/Ca Ratios - Comparison of Lake and Springwater During Precipitation of Gypsum

The salts in the Tyrrell lake water have a large excess of NaCl from halite re-solution, and their initial percentage magnesium content is much lower than that in the springwater at any given salinity or chlorinity (Figure 9.11). Furthermore, the lake waters have a higher calcium content than the western springwaters, giving a lower Mg/Ca ratio prior to gypsum precipitation than is the case in the spring-waters.



Figure 9.15 - Prismatic selenite crystals, showing parallel downward growth within laminated silts of the spring zone. A section containing the crystals has been placed on the ground surface and exposed to weathering to permit the parallel pattern to be more readily seen. The pH of the groundwater from which the selenite is precipitating is about 2.9 to 3.1, and the salinity 300 g/l.



Figure 9.16 – Prismatic selenite crystals

On evaporitic concentration of lake waters, the Mg/Ca ratios range from about 1.2 to 10 during the main gypsum phase, and reach a maximum of about 16 to 18 only during the halite stage, when the S.G. is about 1.24 (Table 10.1). By comparison, waters within the spring zone have Mg/Ca ratios which are around the 16 to 18 mark prior to the commencement of gypsum precipitation, at an S.G. of about 1.17. As gypsum precipitation commences, the Mg/Ca ratio rises to 30 and continues to rise through the gypsum and then the gypsum–halite phases. In the more concentrated solutions, Mg/Ca ratios reach as high as 60 to 80 while the highest S.G. measured in the lower spring zone is 1.225.

9.4.5 Vertical Selenite Crystals—the Messinian of Sicily

Perhaps the best documented occurrence of vertically standing selenite comes from the Solfifera Series of the Messinian of Sicily, where coarse selenite, with single crystals up to one metre in length, has been observed in beds up to 30 m thick. Hardie and Eugster (1971) describe selenite crystals from the Solfifera Series. They consider the selenite to be primary, and that it grew in increments contemporaneously with deposition of the surrounding fine-grained matrix. They comment that 'most of the growth of the selenite crystals must have taken place at or very near the surface, either sub-aerially or under a shallow brine'. They suggest that the most likely depositional environment is that of a shallow lagoon. Schreiber and Kinsman (1975) described beds of elongate vertical gypsum crystals from the Pleistocene Montalegno Formation, derived from the Messinian beds. The selenite crystals are elongated parallel to the 'c' crystallographic axis. Individual crystals are up to 10 cm in length, roughly conical in shape, and increasing in width upward. They suggest that crystal growth is by direct subaqueous precipitation, arising from nucleation at the brine–sediment interface. They suggest a subtidal to intertidal environment. Schreiber et al. (1976) agree with Hardie and Eugster's assessment of the depositional environment for the primary gypsum, but give the continuity of gypsum beds across 12–15 km of basin as evidence of somewhat deeper water.

Vai and Lucchi (1977) also discuss the origin of the Messinian gypsum beds. They consider that the upstanding selenite crystals formed at or below the sediment–water interface within algal mats. The selenite grew concomitantly with algal mat growth and therefore is contemporaneous with sedimentation. In order to provide a source for large accumulations of clastic gypsum found in the Messinian sediments, they suggest that the autochthonous gypsum must come from nearshore and coastal areas. It is thus exposed to erosion and mass transport, with little abrasion of clasts followed by rapid entombment. 'Torrential flash floods emptying into a quiet water body provide the suitable mechanism'. A subtidal to intertidal lagoonal environment is favoured; however, they acknowledge that selenite may also form in supratidal marshes.

9.4.6 Prismatic Groundwater Gypsum - A Comparison

Apart from crystal form and growth positions, there seems little else in common between the depositional environments of the massive marine selenites of Sicily and the thin gypsum bands in the T–P line of the western shoreline of Lake Tyrrell. The only overlap would seem to be that both occurrences are in a shoreline or nearshore environment - in the Tyrrell case, supratidal from acid groundwater and in the Sicilian case, intertidal to subtidal from marine waters. But perhaps the greatest difference is that the Tyrrell gypsum is forming at present, yet is entirely subsurface. By contrast, the Sicilian selenite has grown concurrently with sedimentation, whether it be in an algal mat or carbonate–gypsum matrix. In this respect, the Tyrrell gypsum does not show the striate or banding within the individual crystals, as recorded by Hardie and Eugster (1971) in the Marion Lake deposits of the Yorke Peninsula in South Australia. The laminae on the gypsum from Marion Lake were seen as a record of episodes of contemporaneous dissolution and regrowth prior to the deposition of subsequent layers. Dickinson and King (1949) considered the layering to reflect seasonal variations in lake salinity.

9.4.7 The Influence of pH on Crystal Growth

It has been previously noted that both the presence of organic matter and pH may play major roles in the development of crystal form (Edinger, 1973; Cody, 1976). Cody showed that lenticular gypsum crystals grow in bentonite under alkaline conditions, but elongate crystals form in acid conditions. The elongate crystals were ascribed to the adsorption of H^+ on the {010}, and perhaps {110} faces, allowing growth perpendicular to {111} and {1103}. Later work by Cody (1979) showed that, despite the presence of organic matter, which would normally slow growth perpendicular to the c-axis and thus promote lenticular growth, acid conditions not only favoured growth parallel to the c-axis but also greatly increased the total gypsum yield. Commenting on the geological significance of his work, Cody notes that the presence of elongate prismatic crystals in organic media such as black shales and coal is probably due to low pH effects. He also notes that, although the organic matter present may not be of a type to influence crystal growth, a more plausible conclusion is that low pH is the major factor promoting the growth of prismatic gypsum. The shoreline environment at Lake Tyrrell provides an ideal range of hydrochemical regimes in which to examine the varying forms of gypsum under a number of pH conditions. In this respect, the marked contrast in pH between the eastern and western spring zones is enhanced by the gradual increase in pH across the western spring zone as the lake is approached. The results support Cody in that the discoidal gypsum is being extensively precipitated in the sediments of the eastern spring zone while prismatic crystals are forming from acid solutions in the western spring zone. In the latter case, as the pH of the shallow groundwater rises on passing lakewards, the crystals become increasingly stubby, and the prismatic form virtually disappears (Figure 9.17). It is clear, therefore, that high acidity is the overriding factor in the development of prismatic gypsum crystals in the sediments of the western spring zone of Lake Tyrrell. The Tyrrell data supports Cody's experimentally based conclusions that prismatic gypsum crystals may in certain instances be an indicator of acid conditions at the time of their formation.

9.5 Eastern Spring Zone and Nearshore Lake Floor

The eastern spring zone is fed largely by the Timboram–Wahpool reflux brines, and consequently springhead waters rarely have salinities less than about 100,000 mg/l (Table 9.5). Unlike the situation on the western shoreline, there are relatively few bores on the eastern shoreline. Some idea of the vertical distribution of salinities can be obtained from the SR 84 piezometer nest on the northeastern lakeshore (for location, see Figure 7.5a), where bores at depths of 1.85 m (SR 84A), 5 m (SR 84), and 20.8 m (SR 84 B) have salinities of 97,400, 146,200, and 163,800 mg/l respectively (Figure 9.18). While the salinity of the uppermost groundwater is close to that of the west-flowing Timboram–Wahpool reflux brines, the deeper waters have salinities intermediate between these brines, and those of the Tyrrell reflux brines which underlie Lake Tyrrell, and are present at depth below the piezometer nest. Thus, bores 84 and 84B are seen as lying in the transition zone between the two reflux brines; both bores show the higher Mg/Br ratios characteristic of the Timboram-Wahpool brines.



Figure 9.17 - Stubby and irregular gypsum, precipitated from groundwater in the zone of overlapping spring and lake waters lakeward of that shown in Figure 9.15. The loss of prismatic morphology reflects the rising pH under the influence of the neutral pH lake regime.

Prismatic and discoidal gypsum crystals are included for comparison.

As is the case with similar nests on the western shoreline of the lake, the upward movement of groundwater on the eastern shoreline is shown by an upward potentiometric gradient; for on passing from deepest to shallowest the static levels are 43.78, 43.25, and 42.84 m AHD respectively. It is notable that all three static levels are well above the highest levels normally reached by the lake (about 42 m AHD). While the 84 nest did not penetrate the Tyrrell reflux brines, these brines were encountered in the deeper Lianiduck 10002 bore, situated 2 km to the north of the 84 nest.

Apart from the springs and bores situated along the eastern shoreline, the character of the inflowing groundwaters and their modification in the near surface sediments was investigated by an examination of waters from two lines of piezometers extending out across the lake floor from the shoreline. The piezometers are shallow, being about 1 m deep. The shallow groundwaters are influenced by surface processes and reflect both the chemistry of the upwelling reflux brine and its modification arising from lake-groundwater interactions.

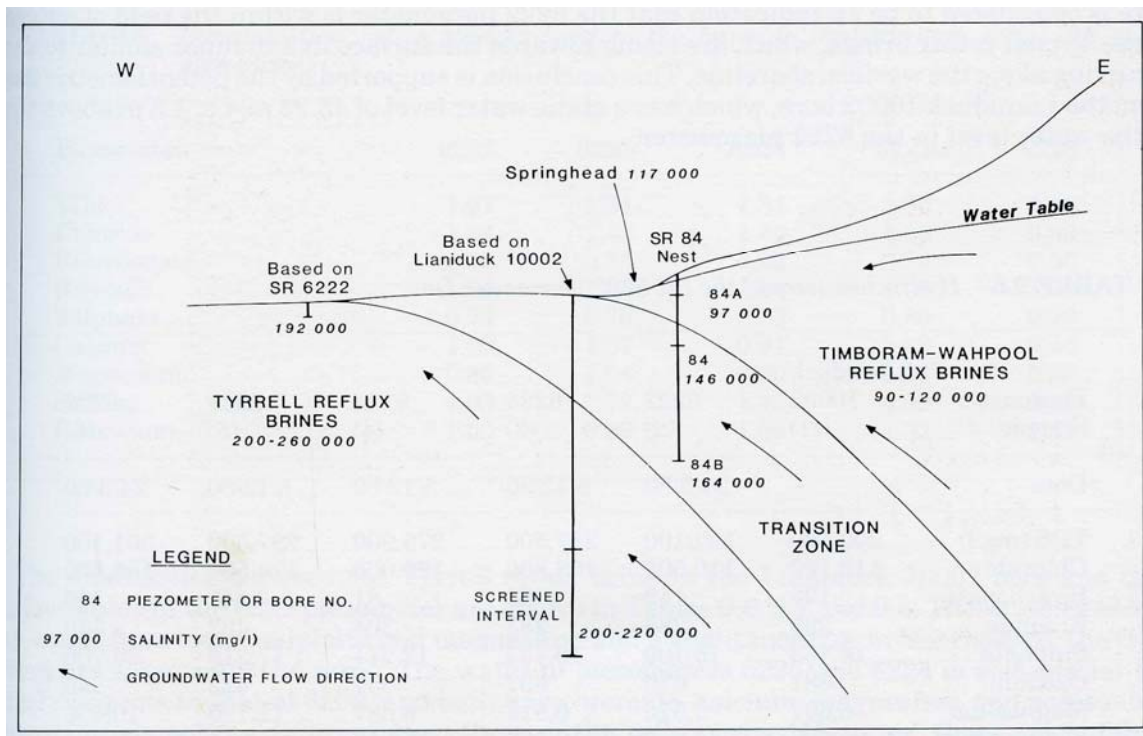


Figure 9.18 - Hydrochemical zonation across the eastern shoreline of Lake Tyrrell. This figure is diagrammatic and is a composite of piezometer data from along the shoreline. The section shows the broad transition zone between the Tyrrell reflux brines and the Timboram-Wahpool reflux brines.

The clearest picture is obtained from the SR 6222-6225 line of four piezometers, extending lakewards from the southeastern shoreline. The salinities range from about 190,000 mg/l in piezometer 6222, closest to the shoreline, to about 300,000 mg/l in piezometer 6223, which although not the furthest lakewards occupies the lowest point on the line. The high salinity of the piezometers is well beyond the range of the Timboram-Wahpool reflux brine, but comes close to that of the Tyrrell reflux brine (analysis 1, Table 9.6) as found at a depth of 24-47 m in the Lianiduck 10002 bore, situated on a peninsula immediately lakewards of the shoreline, but well to the north of the piezometer line. This is the deepest bore on the eastern shoreline and the only eastern bore to reach the underlying Tyrrell reflux brine.

The bore water is compared with the shallow piezometer waters in Table 9.6. The marked similarity between the 6222 water and that of the Tyrrell reflux brine in the Lianiduck 10002 bore is considered to be an indication that the 6222 piezometer is within the field of outcrop of the Tyrrell reflux brines, which are rising towards the surface in a manner similar to that occurring along the western shoreline. This conclusion is supported by the potentiometric data from the Lianiduck 10002 bore, which has a static water level of 43.23 m, i.e. 1.5 m above that of the water level in the 6222 piezometer.

TABLE 9.6 - Hydrochemistry of the SR 6222 piezometer line

Piezometer Sample	Lianiduck					
	10002 (1)	6222 (2)	6225 (3)	6224 (4)	6223 (5)	6223 (6)
Date	-	24/3/80	8/12/80	8/12/80	8/12/80	24/3/80
TDS (mg/l)	200,600	192,100	287,500	275,200	297,500	301,100
Chloride	112,100	110,500	158,800	160,000	170,500	168,500
Bicarbonate	128	143	64	81	141	140
Bromide	299	285	283	304	422	540
Sulphate	14,483	10,100	10,433	12,274	17,219	20,700
Calcium	681	670	697	657	422	540
Magnesium	6,607	5,370	6,517	6,087	11,110	10,400
Sodium	65,723	64,500	110,200	95,260	96,600	99,200
Potassium	526	500	496	570	1,025	1,115
Silica	5	6	4	6	3	3

While the shallow piezometers show the influence of the upwelling Tyrrell reflux brines, it is notable that at the springhead on the shoreline to the east of 6222 the salinity (117,300 mg/l) and overall chemistry closely resemble that of the Timboram-Wahpool reflux brine (see analysis 4, Table 9.5). Although not closely examined, it is assumed that a range of salinities exists between the springhead and the 6222 piezometer, which reflects the outcrop of the transition zone between the Timboram-Wahpool and Tyrrell reflux brines in a manner comparable to that examined at the T-P line on the western shoreline (Figure 9.18).

TABLE 9.7 - Concentration ratios relative to each element of the Lianiduck 10002 bore

Piezometer Sample	Lianiduck					
	10002 (1)	6222 (2)	6225 (3)	6224 (4)	6223 (5)	6223 (6)
TDS (mg/l)	1	0.96	1.43	1.37	1.48	1.50
Chloride	1	0.99	1.42	1.43	1.52	1.50
Bicarbonate	1	1.12	0.50	0.63	1.10	1.09
Bromide	1	0.95	0.95	1.02	1.48	1.81
Sulphate	1	0.70	0.72	0.85	1.19	1.43
Calcium	1	0.98	1.02	0.96	0.62	0.79
Magnesium	1	0.81	0.99	0.92	1.68	1.57
Sodium	1	0.98	1.68	1.45	1.47	1.51
Potassium	1	0.95	0.94	1.08	1.95	2.12

TABLE 9.8 - Concentration ratios standardized to the bromide concentration ratios for each piezometer

Piezometer	6222	6225	6224	6223	6223
TDS	1.01	1.51	1.34	1.00	0.83
Chloride	1.04	1.49	1.40	1.03	0.83
Bicarbonate	1.18	0.53	0.62	0.74	0.60
Bromide	1.00	1.00	1.00	1.00	1.00
Sulphate	0.74	0.76	0.83	0.80	0.79
Calcium	1.03	1.07	0.94	0.42	0.44
Magnesium	0.85	1.04	0.90	1.14	0.88
Sodium	1.03	1.77	1.42	0.99	0.83
Potassium	1.00	0.99	1.06	1.32	1.17

The similarities between the Tyrrell reflux brine in the Lianiduck 10002 bore and the shallow

water in the 6222 piezometer are shown in Tables 9.6, 9.7 and 9.8. These tables show that, apart from lower sulphate and magnesium and a slight increase in bicarbonate, the two waters are essentially the same. The water in piezometers 6225 and 6224 is also similar in certain respects to that of 6222, and both have bromide, calcium, magnesium and potassium concentrations not too different from those of the reflux brine. However, there are notable salinity differences (which can be attributed, almost entirely, to the gain of sodium chloride by water) in the 6225 and 6224 piezometers. Halite pickup-shown by higher Cl/Br ratios (Table 9.10)-occurs in piezometers established beyond the influence of the upper spring zone, where positive hydrostatic pressures, coupled with continuous outseepage, enable the chemistry of the upwelling reflux brines (as in piezometer 6222) to remain comparatively unaltered. Lakewards, where the shallow groundwaters come under the influence of halite-enriched lake floor regimes, they are saturated with halite and thus have higher Cl/Br ratios (as shown in the analyses from 6223, 6224 and 6225, Table 9.6).

The process of halite enrichment in these shallow groundwaters is shown below as being essentially a seasonal process, commencing from the time when the lake begins to dry out in summer. Two processes contribute to the enrichment of shallow groundwaters under summer regimes: (1) in the final stages of halite deposition as the lake dries out, the water table falls beneath the salt crust and carries the bitterns from the lake regime into the shallow groundwater regime; (2) once the lake floor is dry, the water table begins to oscillate within the very shallow subsurface regime and, in doing so, redissolves some of the surface halite crust until halite saturation occurs or until the salt crust is removed. These latter processes are discussed in detail in Chapter 11. They are hydrodynamic processes during which the water table moves vertically through the sediments and the salt crust.

By contrast, recharge of the shallow subsurface sediments under lake-full conditions is essentially a hydrostatic process whereby lake water migrates down through the lake floor under a downward hydraulic gradient. From Table 9.9, it can be seen that while the lake waters have a high Cl- content they are comparatively low in Br-, resulting in Cl/Br ratios of about 1000. It is only after halite precipitates in the later stages of evaporitic concentration (Table 9.9, analysis 12) that the Cl/Br ratio of the lake waters approaches the values found in the piezometers (Table 9.10). Therefore, while the reflux of lake waters by hydrostatic processes under lake-full conditions would raise the chloride content of the shallow groundwaters, it would also impart high Cl/Br ratios such as those shown for lake waters in Table 9.9, analyses 7 to 11.

TABLE 9.9 - Chloride-bromide characteristics of the Tyrrell lake waters (mg/l)

Sample Date sampled	(7) 25/5/80	(8) 30/5/80	(9) 16/7/80	(10) 26/4/80	(11) 14/10/80	(12) 10/11/80
TDS	290,377	285,482	284,174	310,518	323,887	339,926
Chloride	168,750	167,000	165,700	183,310	190,800	186,000
Bromide	155	175	170	150	165	351
Cl/Br	1,089	964	975	1,222	1,162	530

TABLE 9.10 - Chloride-bromide characteristics of the 6222 line (mg/l)

Piezometer	Lianiduck 10002	6222	6225	6224	6223	6223
Salinity	200,561	192,079	287,500	275,241	297,464	301,145
Chloride	112,100	110,500	158,800	160,000	170,500	168,500
Bromide	299	285	283	304	442	540
Cl/Br	375	388	561	526	386	312

This is the situation when the lake waters are driven by the wind to invade either the lower spring

zone or areas of the lake floor previously dry. Examples of this process are given in Table 9.3, where the Cl/Br ratio jumped from 286 to 909 in the shallow (0.5 m deep) piezometer 1 between April and May 1981 - the higher ratios are those of the infiltrated lake water, while the lower ratios are typical of the outflowing springwaters. Despite the existence of high Cl/Br groundwaters in the uppermost parts of the aquifer, these waters are not otherwise found within the aquifer and their presence is a temporary seasonal phenomenon.

The samples in Table 9.6 were obtained in summer and early autumn, when the water tables are oscillating through the salt crust. It is likely that the processes reflected by the shallow groundwater chemistry at the sites of piezometers 6224 and 6225 are due to hydrodynamic processes active during the final stages of lake drying and after the lake has dried out. The highest concentrations of the bittern ions, bromide, potassium, magnesium and sulphate, occur in piezometer 6223 (Table 9.6). The presence of the high bittern component is seen largely as a reflection of the position of this piezometer which is at the lowest point on the piezometer line, a position on the lake floor where the final bittern enriched remnants of lake water undergo reflux processes as the water table falls below the lake floor.

In conclusion, therefore, while the hydrochemistry of the 6222 piezometer line shows the underlying chemistry of inflowing reflux brines, the temporal and spatial variations in ionic concentrations in the piezometers show the influence of additional processes reflecting the interactions between the lake system and the shallow groundwater system.

10 HYDROCHEMISTRY OF LAKE WATERS IN NORTHWESTERN VICTORIA AND THE INFLUENCE OF REGIONAL GROUNDWATERS

10.1 General

The Mallee region of northwestern Victoria is essentially an aeolian landscape, with intrusion of riverine systems only along the Wimmera River, and to a very minor extent along the Yarriambiack, Tyrrell and Lalbert Creeks; all four streams peter out well before they reach the Murray River. With the absence of any significant riverine contribution, the surface waters seen in the many ephemeral lakes within the Mallee are derived from direct precipitation, localized runoff, and from groundwater sources. It became clear during the early stages of the investigation that there was a significant degree of interaction between the lake systems and the groundwater systems, which was strongly reflected in the hydrochemistry of the lakes.

10.2 Hydrochemistry of Lake Tyrrell

At times, Lake Tyrrell (Figure 10.1) is completely dry. At other times, a large part of the floor is covered by several tens of centimetres of water (Figure 10.2). No long-term record of depth variation has been kept, but maximum water depths within historic times probably have not exceeded one metre. Because of the variable volume of water in the lake, its chemical composition varies, although it always remains highly saline. The chemistry of the Tyrrell lake waters at any particular time is the resultant of halite re-solution and precipitation processes as the lake passes through its annual cycle of wetting and drying. In winter, the salinity fluctuates between about 250,000 mg/l and 300,000 mg/l; however, in early summer it reaches 330,000 mg/l as evaporitic concentration takes place. After complete evaporation, some 7 cm of halite covers the lake floor (Figure 10.3), only to be redissolved during the following wet season. Because the bulk of the salt entering the lake is of groundwater origin, the chemistry of the lake waters is very similar to that of waters in the regional Parilla Sand aquifer. However, once in the lake, the waters show marked changes in ionic ratios, as the result of seasonal halite re-solution followed by evaporitic concentration; during the latter stage gypsum and halite are precipitated (Table 10.1).

The samples in Table 10.1 are actual lake samples or, in the case of the more concentrated waters, residual brines from beneath the salt crust. They therefore reflect not only the general trends in chemical evolution of the lake waters during their annual cycle of re-solution and evaporitic concentration, but also the variations produced by groundwater additions and brine reflux, and by the movement of lake waters from place to place by wind action. These factors combine to produce the deviations from a theoretically derived chemical evolution sequence.

TABLE 10.1 - Evaporitic concentration during drying of Lake Tyrrell⁽¹⁾

	S.G.	Ca ²⁺	Mg ²⁺	Na ⁺	SO ₄ ²⁻	Cl ⁻	Cl/Mg	Cl/Na	Cl/SO ₄	Cl/Ca
1.	1.16	1457	2888	83200	11000	130800	45.3	1.57	11.9	90
2.	1.20	720	3058	115400	7440	183100	59.9	1.59	17.9	254
3.	1.21	874	8900	117200	8950	187600	56.8	1.60	21.1	214
4.	1.21	951	4115	117600	10040	187700	45.6	1.60	18.7	197
5.	1.215	1108	5720	112900	9480	185800	32.5	1.65	19.6	168
6.	1.22	791	9740	112400	20860	184300	18.9	1.64	8.8	233
7.	1.23	923	12321	105300	30750	178100	14.5	1.69	5.8	193
8.	1.23	1190	16580	96800	35180	174300	10.5	1.80	5.0	146
9.	1.245	1520	27170	94100	63850	180600	6.7	1.92	2.8	119
10.	1.26	1930	32180	81800	78250	166700	5.2	2.04	2.1	86
11.	1.03 ⁽²⁾	470	1350	11070	3190	19800	14.6	1.78	6.2	42
12.	1.025 ⁽³⁾	400	1300	10500	2650	19000	14.9	1.80	7.2	47
13.	1.16 ⁽⁴⁾	514	7100	77159	15452	131000	18.4	1.70	8.5	254

- (1) Table based partly on data supplied by Cheetham Salt; halite precipitation commences at S.G. 1.21; note CaSO₄ decline between S.G. 1.16 and 1.215 as gypsum precipitates
(2) Regional Parilla Sand water
(3) Seawater
(4) Tyrrell reflux brine

In some respects, therefore, the lake waters are significantly different from the parent groundwaters (Table 10.1, analysis 11). Such differences are especially evident in winter, when the lake is at its lowest salinity and lowest density. At this stage, the NaCl component in the lake waters far exceeds that in the groundwaters, demonstrating the strong influence of halite re-solution on the lake chemistry. Apart from the excess NaCl, the lake waters resemble a concentrated groundwater that has reached a salinity of about 110,000 mg/l and then gained a further 170,000 mg/l to 220,000 mg/l by halite re-solution. Indeed, if the excess halite were to be removed, the Tyrrell waters would resemble the Timbora-Wahpool reflux brines (see Table 10.2).

These analyses show clearly the similarity in ionic concentrations between the 28+‰ lake waters and the 11‰ Timboram reflux brines for all but the sodium, chloride and silica ions. Conversely, while the Tyrrell reflux brines have similar salinity, sodium and chloride concentrations to the lake waters, they differ greatly in their respective bittern ion concentrations (i.e. Mg^{2+} , Br^- , K^+ , and SO_4^{2-}). These differences are important when considering the origins of the Tyrrell reflux brines, since they indicate that the brines were not formed by the simple infiltration or reflux of lake water which had characteristics similar to those of the present lake. The relationship of the lake water chemistry to that of other waters in the Tyrrell Basin can be readily seen in Figures 7.2 and 10.4. Figure 7.2 shows the magnesium versus bromide plot for all surface waters and groundwaters in the Tyrrell Basin, their salinities ranging from 30,000 to 350,000 mg/l. The graph shows the Mg/Br ratio to be fairly constant with increasing concentration. The Tyrrell lake waters fall on the line part way along the curve, in accord with the degree of concentration of their regional groundwater component.

TABLE 10.2 - Comparison of the Tyrrell lake waters with the Timboram—Wahpool and Tyrrell reflux brines

	Tyrrell reflux brine	Timoram- Wahpool reflux brine	1	Tyrrell lake waters 2	3
Date sampled			16/7/80	30/5/80	20/10/80
Total salinity	267,789	114,308	274,174	285,482	324,497
Sodium	87,800	35,460	104,300	103,600	119,000
Chloride	150,750	64,642	165,700	167,000	189,300
Sulphate	18,150	8,023	9,520	9,800	10,800
Bromide	455	176	170	175	194
Calcium	550	815	770	875	883
Magnesium	9,250	4,735	3,300	3,300	3,850
Potassium	810	333	363	375	400
Silica	11	13	2	3	8
Cl/Na	1.72	1.82	1.61	1.61	1.59
Cl/Mg	16.3	13.6	46.7	46.7	49.2
Cl/ SO_4	8.3	8.1	18.1	17.0	17.5
Cl/K	186	194	445	445	478
Cl/Br	331	367	982	954	978

Apart from the work of McLaughlin (1966), the only other study of Victorian lake waters which allows a comparison of Mg/Br ratios is that of Maddocks (1967), who analyzed surface waters from lakes, maars, and streams in southwestern Victoria. Maddocks's analyses are specially useful for comparative purposes in that they have a wider salinity range than that which occurs in the more arid northwestern Victoria. As is the case with the Mallee waters, Maddocks's waters have a distinct oceanic character, which may be seen in the Cl/Br and Na/1 ratios (see also Chapter 7.4). In Maddocks's fresher waters, magnesium, expressed as a percentage of the total cations, shows a distinct decline as the salinity rises; he attributes this loss to the precipitation of dolomite. More detailed examination of this drift (Table 7.5) shows that the major magnesium loss is in waters having a salinity of less than 5,000 mg/l; there is no significant difference in the waters ranging from 5,000 to 100,000 mg/l. Although a further fall in magnesium content in waters over 100,000 mg/l is suggested, the significance of this is difficult to assess given the small number of samples (three) and the large standard deviation, although if it is real, it may be due to sepiolite formation (cf. McLaughlin, 1966).

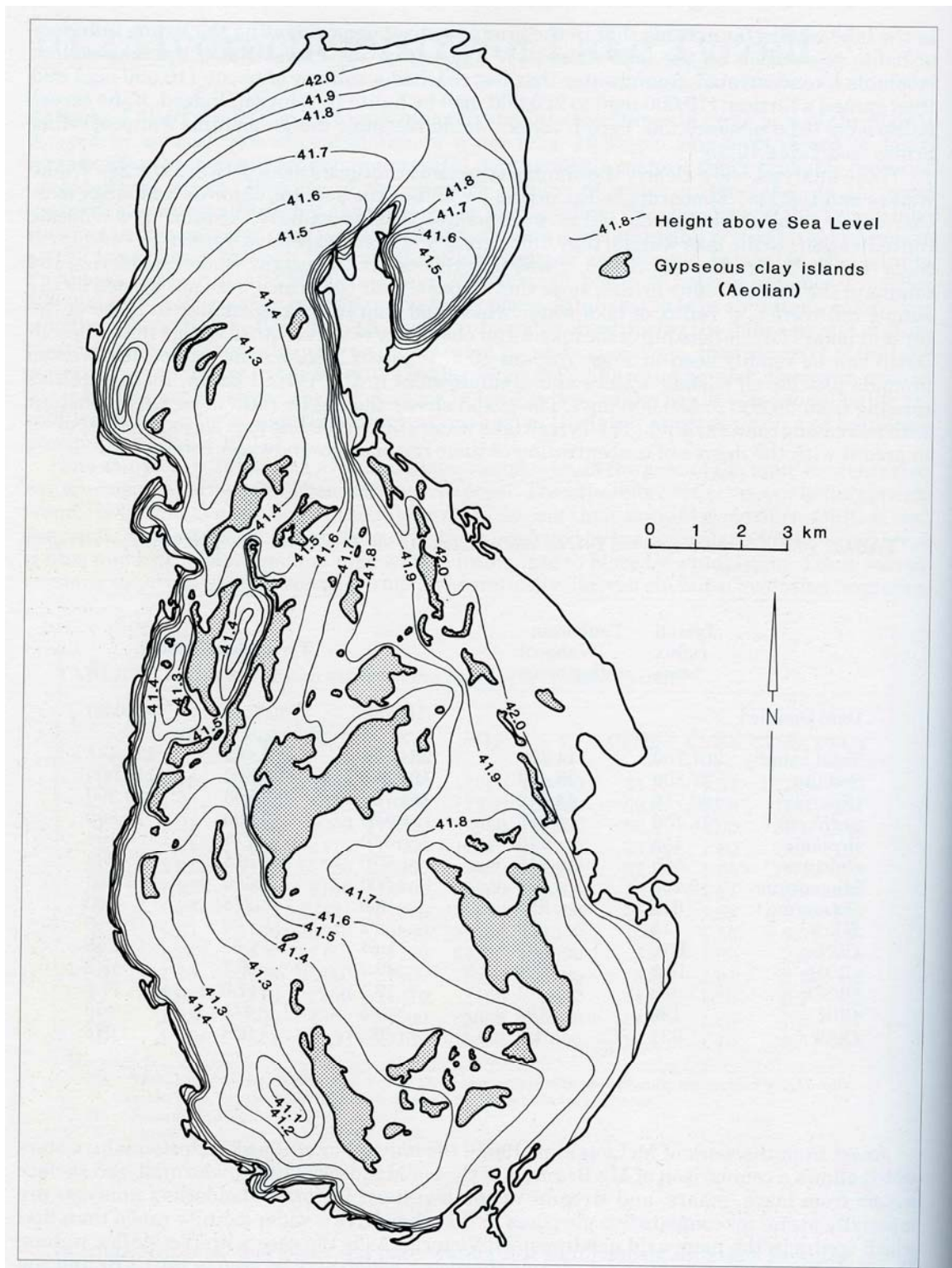


Figure10.1 - Contour map of Lake Tyrell



Figure 10.2a - Lake Tyrrell, looking north. The lake has a shallow water cover and gains its pinkish tint from the presence of *Dunaliella salina* algae. The lake is about 26 km long and 8 km wide. A number of aeolian gypseous clay islands are scattered across the lake floor; in the lower right is the Tyrrell Creek, an essentially base flow stream with a pH of 4.0.



Figure 10.2b – Lake Tyrrell in winter – the water depth is about 0.4 m and its salinity about 290 g/l



Figure 10.3 - Salt crust on the floor of Lake Tyrrell. In summer the crust is about 7 cm thick, but it redissolves in winter to give a NaCl enriched lake water.

Despite the loss of magnesium in the more dilute range, once the waters have salinities in excess of 5,000 mg/l there is an almost identical relationship between bromide and magnesium as that which occurs in the lakes and groundwaters of northwestern Victoria (see also Chapter 7.4). The bromide versus chloride plot of the Tyrrell Basin waters (Figure 10.4) gives a similar concentration curve, but here the lake waters are well off to the chloride side of the curve, reflecting the effects of the excess halite gained by re-solution of the annual halite crust. By contrast, all the concentrated groundwaters, including the Tyrrell reflux brines, lie directly on the curve, showing the absence of a comparable re-solution phase during their evolution. One notable effect of the halite re-solution is that the chloride - bromide ratios have been pushed from about 350 to over 950 (Table 10.6, analysis 11).

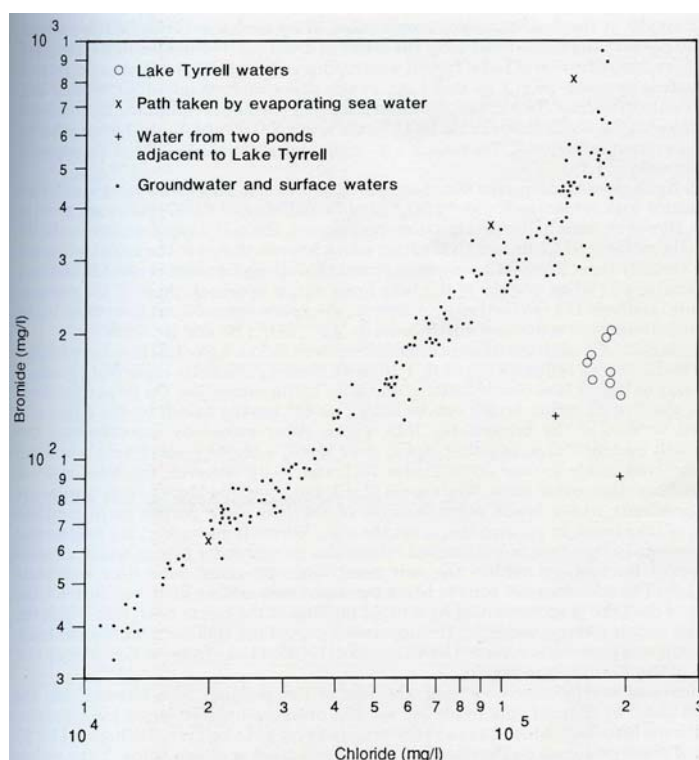


Figure 10.4 - Log bromide vs. log chloride for all surface waters and groundwaters in the Tyrrell Basin (seawater data from Bassegio, 1976)

10.3 Evaporitic Concentration of Lake Tyrrell Waters

Whenever evaporitic concentration commences in Lake Tyrrell, the ionic ratios undergo significant changes as firstly gypsum then halite are precipitated. As a consequence, no single analysis adequately describes the chemical character of the evaporitic lakes, but instead a number of samples must be taken over the annual cycle of re-solution and concentration.

The figures given in Table 10.1 cover the entire range of salinities of the lake waters which are normally encountered. The very highest densities shown in the table are not commonly met, except maybe at the final stages of evaporation. They are more typical of the residual waters found beneath the halite crust after the lake has dried out. During the initial stages of the evaporative concentration of Lake Tyrrell waters, there is a notable fall in the calcium and sulphate content between S.G. 1.16 and 1.20; at the same time, magnesium, sodium and chloride concentrations rise. This covers the main phase of gypsum precipitation. It is reflected in the sharp rise in the Cl/Ca ratio (Table 10.1). In the lower S.G. range, the influence of halite re-solution is evident in the low Cl/Na ratio of 1.57 (close to that of halite - 1.54, cf. the regional groundwater ratio of 1.78).

The high NaCl content of the lake water compared to the regional groundwater is reflected in high Cl⁻ ratios with respect to K⁺, Mg²⁺, SO₄²⁻, and Br⁻ (Cl/Mg and Cl/SO₄ ratios are given in Table 10.1). However, once halite precipitation commences, there is a rapid decline in the Cl/SO₄ and Cl/Mg ratios (and Cl/Br and Cl/K ratios which are not shown in the table), while the Cl/Na ratio steadily rises. The NaCl content continues to fall, and a stage is reached at about S.G. 1.23 (analysis 7) when certain of the lake brine ratios approach those of the regional groundwaters (analysis 11). On further evaporation, the ratios again depart from those in the regional groundwaters as continued enrichment in Mg²⁺, SO₄²⁻, K⁺ and Br⁻ occurs.

It is notable that, although the concentrated lake waters at S.G. 1.22–1.23 may have similar ionic ratios to the Tyrrell reflux brines (S.G. 1.16) with their significantly lower NaCl content, this occurs only at higher lake concentrations and after halite saturation. On these considerations alone, the Tyrrell reflux brines can be shown as not having arisen by the reflux of a concentrated version of the present-day lake water. After evaporitic concentration, the magnesium salt content (taken as either MgCl₂ or as MgCl₂ + MgSO₄) rises from about 6% of the total dissolved solids to over 30% (Tables 10.1 and 10.3); however, the lake does not normally produce magnesium salts. While some MgCl₂ (and perhaps MgSO₄) may precipitate in local depressions where lower permeabilities of the lake floor permit more complete evaporation of lake brine, in general this is not the case. Where it does occur, the magnesium salts are soon redissolved by undersaturated interstitial groundwater brines, which undergo diurnal vertical fluctuations within the salt crust and uppermost lake floor sediments (Chapter 11.8). The lake does not behave like a porcelain evaporating dish, but, instead, the final drying of the lake is accompanied by a rapid buildup of the halite crust into which the remnant lake waters retreat, becoming the uppermost part of the shallow groundwater body. A similar result was previously recorded by Valyashko (1968) at Lake Inder in Kazakhstan and Lake Elton in the Transvolgan region.

In the lakes of northwestern Victoria, the loss of the residual lake bitters into the sediments is aided by diurnal and, in places, weekly, pressure-induced water table fluctuations beneath the lake floor, which have a range of up to 24 cm at Lake Tyrrell (Chapter 11.2.2). The effects of these processes on the chemistry of the salt crust is shown below. Lake waters (Table 10.3), their composition showing the seasonal cycle of evaporitic concentration, are compared with salt crust analyses (Table 10.4) and with interstitial brines from beneath the salt crust (Table 10.5).

Although the MgCl₂ content of the lake waters during the late stage of evaporation is 28.4% (sample 4), the actual amount of MgCl₂ retained in the salt crust is only 0.5%, while varying amounts up to 28% were found in the subsurface brines. While it might be initially considered that the brines should all contain a similarly high MgCl₂ content to that in sample 4, this is not the

case, since the residual lake waters while passing into the groundwater system mix to varying degrees with the pre-existing groundwaters which have a lower MgCl_2 content. In this example the Mg^{2+} is taken as being MgCl_2 ; however, even if the Mg^{2+} is allocated to SO_4^{2-} as well as Cl^- , it does not alter the percentage of magnesium salt in the waters. The same processes occur in the case of Na_2SO_4 , which reaches a maximum concentration of 8.5% in the concentrated lake waters, but is only 0.4% in the salt crust, and generally less than 0.2%. In the groundwater brines it varies from about 6% to 8.5%, reaching 12.8% in one sample. The lower solubility of gypsum in these saline waters is reflected both in the relatively low CaSO_4 content of the lake waters and in the higher CaSO_4 content of the salt crust than that in the groundwater brines.

TABLE 10.3 - Lake water from southwestern Lake Tyrrell-salt content

Sample no. Date sampled	1 30/5/80 %	2 16/7/80 %	3 10/11/80 %	4 2/12/80 %	5 19/8/81 %
NaCl	89.5	90.1	83.0	62.1	92.6
Na_2SO_4	3.3	3.3	5.5	8.5	2.5
MgCl_2	6.1	5.6	10.7	28.4	4.1
CaSO_4	1.9	0.8	0.5	0.2	0.8

TABLE 10.4 - Salt crust analyses-SW Lake Tyrrell (sampled on 6/2/81)

Sample no.	6 %	7 %	8 %	9 %
NaCl	98.7	98.7	98.6	99.0
Na_2SO_4	0.4	0.1	0.1	-
MgCl_2	0.5	0.5	0.5	0.3
CaSO_4	0.4	0.7	0.7	0.7

TABLE 10.5 Salt content of brine from beneath the salt crust

Sample No.	SW Lake Tyrrell Sampled February 1981				T -P1 piezo NW Tyrrell Jan. 1981	6223 piezo SE Tyrrell Dec. 1980
	10 %	11 %	12 %	13 %	14 %	15 %
NaCl	68.2	71.9	81.6	83.4	58.6	74.8
Na_2SO_4	6.4	8.5	6.2	6.4	12.8	7.8
MgCl_2	24.6	18.8	11.5	9.6	27.8	16.4
CaSO_4	0.3	0.4	0.4	0.5	0.2	0.5
KCl	0.3	0.5	0.3	0.3	0.6	0.5

10.4 The Bromide Versus Chloride Framework for Playa Lakes of Northwestern Victoria

Lake Tyrrell is one of a relatively small number of re-solution lakes from northern Victoria. The waters of these lakes are almost always saturated with halite which is principally derived by re-solution of a seasonally precipitated halite crust. Since the lake water is virtually saturated with NaCl at the commencement of the evaporative concentration stage, the bromide versus chloride graph of the evaporating lake waters shows a near vertical plot which is parallel to the bromide axis (Figure 10.4). By contrast, playa lakes in which there is little or no seasonal accumulation of halite have a water chemistry closely mirroring the regional Parilla Sand groundwaters (see Lake Timboram, following). On evaporative concentration, therefore, the Br/Cl plot follows a concentration curve which is essentially the same as that for concentrating seawater; it parallels the bromide axis only after halite saturation is reached. In lakes where the initial stage of halite re-solution does not lead to halite saturation prior to the commencement of the evaporative phase, the plot of bromide versus chloride commences between the seawater concentration curve and the halite saturation zone. As evaporation proceeds, the plot parallels the seawater curve until halite saturation is reached, at which point it turns to parallel the bromide axis.

The bromide versus chloride framework (Figure 10.5) provides a simple technique for categorizing salt lakes across northwestern Victoria. A number of examples are given in Figure 10.6. The simplicity of the system derives largely from the uniform chemistry of the regional groundwaters feeding the discharge lakes. Since the accumulation of halite in a playa lake is largely dependent upon its degree of closure to outseepage (Chapter 12.2), the bromide versus chloride categorization provides an indication of lake hydraulics as well as lake chemistry.

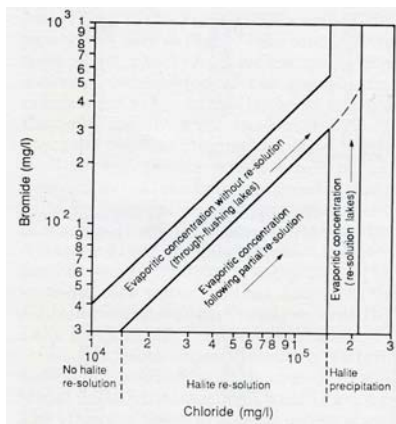
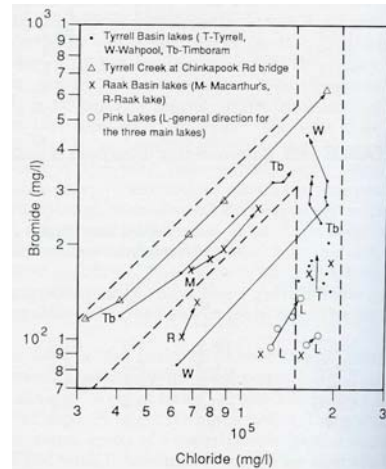


Figure 10.5 - Log bromide versus log chloride provinces for evaporitic regimes in NW Victorian lakes.

Figure 10.6 - The paths followed during the evaporitic concentration of various lake waters from NW Victoria.



10.5 Lake Timboram

10.5.1 Surface Hydrology

Lake Timboram, the terminal lake on the Lalbert Creek (Figure 10.7), is physiographically very similar to Lake Tyrrell, albeit much smaller. The Lalbert Creek flows strongly into the lake only about once or twice each decade, but this is sufficient to put 3 to 4 metres of brackish water over the lake floor. During the abnormally wet years 1973-75, the lake held 3.4 m of water with a salinity sufficiently low to support freshwater fish (Redfin or English Perch) for several years. The upper salinity limit for these fish is about 10,000 mg/l. By October 1978 the lake salinity had risen to only 72,000 mg/l (analysis 1, Table 10.6), but was 280,000 mg/l by May 1980, when the water was restricted to three small sub-basins in the lake floor (analyses 6-8, Table 10.6).

Direct runoff is limited to small catchment areas immediately surrounding the lake, most of which occur in the clayey lunette areas to the east and to the south. Direct precipitation and groundwater inflow make up the remainder of the salt accessions to the lake.



Figure 10.7 - Lake Timboram when dry, with the Lalbert Creek salt flats in the south (bottom). Salt on the lake floor is mostly efflorescing salt, derived by capillary processes from the water table, however, a thin evaporite halite layer is visible in a small depression in the far northern corner of the lake. When Lake Timboram wets up, the halite crust is not present in sufficient quantities to significantly influence the lake chemistry.



Figure 10.8 - Gypsum deposited on vegetation growing around the edge of Lake Timboram. The halophytes were submerged during the 1973-75 wet phase and covered with gypsum during the subsequent drying of the lake.

TABLE 10.6a - Water analyses—Lake Timboram

	TDS	S.G.	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	Cl ⁻	Br ⁻	SO ₄ ²⁻	SiO ₂
1	72000	1.05	1630	2225	-	-	34750	-	6060	-
2	68590	1.05	1448	2280	22500	280	40100	117	1800	0.8
3	108450	1.08	1350	4240	36000	400	63600	159	2560	5.5
4	238140	1.16	980	7740	78800	850	138500	315	10750	7.0
5	244750	1.17	760	8435	79900	825	141000	345	13200	3.0
6	281102	1.19	1150	6600	96300	715	163500	265	12400	7.0
7	293940	1.19	950	8050	100200	835	169000	325	14400	10.0
8	305500	1.20	940	4590	110000	555	179500	230	9630	8.0
9	114308	1.08	815	4735	35460	333	64642	176	8023	13.0
10	36327	1.03	470	1357	11077	113	19757	74	3188	56.0
11	290300	1.19	910	3240	107500	305	168750	155	9450	19.0

Samples 2–3 From McLaughlin (1966)—sampled in winter 1966
 Samples 4–8 Sampled in winter 1980
 Sample 9 Timboram-Wahpool reflux brine (from Table 8.5) Sample 10 Regional Parilla Sand groundwater
 Sample 11 Tyrrell lake water - May 1980

When dry, the Timboram lake floor has at best a thin localized halite crust (Fig 10.7), the re-resolution of which has little influence on the lake water chemistry under lake-full conditions. This situation is in marked contrast to that existing at Lake Tyrrell, where redissolved halite dominates the lake hydrochemistry by making up the bulk of the dissolved salts. As a consequence, the Cl/Br ratio for Timboram is normally only about 400 compared with values of about 1,000 for Tyrrell (Table 10.6). Indeed, the lake-full analyses (1, 2 and 3 of Table 10.6) are more closely akin to the inflowing regional groundwaters (analysis 10), as demonstrated by the ionic ratios. The Cl/Br ratios of Timboram show highest values only in early winter, when the first waters collect in hollows in the lake floor and redissolve the previous year's halite (analyses 6, 7 and 8). Following re-resolution of the halite, further wetting leads to the gradual evolution towards a chemistry which is more typical of the inflowing regional groundwaters (analyses 4 and 5); salinities of less than 100,000 mg/l occur during phases of high lake level in very wet years as shown in analyses 1, 2 and 3.

10.5.2 Hydrochemistry

The path taken by Lake Timboram during evaporitic concentration following the addition of a meaningful volume of water is given by the longer of the two paths marked "Tb" in Figure 10.6. This course follows the straight evaporitic concentration path, with halite re-resolution playing an insignificant role. The shorter path shows the chemical evolution of the lake when it exists only as small isolated pools - here the re-resolution character is brought out, and the path taken by the saturated brine is parallel to the bromide axis.

TABLE 10.6B - Ionic ratios of Timboram lake waters from Table 10.6a

Sample	TDS	Cl/Br	Cl/Na	Cl/Ca
2	68590	342	1.78	28
3	108450	400	1.77	47
4	238140	440	1.76	141
5	244750	409	1.76	182
6	281102	617	1.70	142
7	293940	520	1.69	178
8	305500	780	1.63	191
9	114308	367	1.82	79
10	36327	307	1.78	42
11	290300	1088	1.61	191

In summary, Lake Timboram fills during the occasional flowing of Lalbert Creek. It retains this floodwater for several years, while gradually increasing in salinity. On wetting up after seasonal drying, the lake water rapidly develops a chemical character similar to that of the inflowing regional groundwaters, but quite distinct from the nearby Lake Tyrrell waters with their high Cl/Br ratios from redissolved halite.

10.5.3 Influence of Groundwater–Lake Interactions on Lake Hydro-chemistry

The small amount of halite in the Timboram lake basin reflects the ready interaction between the lake and the groundwater system. Since there is no groundwater divide on the western (downbasin) side of Timboram (Chapter 12.11; Figure 12.1), there is no impediment to groundwater throughflow. Water enters the lake from upbasin and exits through the lake floor on the western side of the lake. During normal seasonal wetting, evaporation accounts for most of the water loss from the lake; however, outseepage becomes significant during each phase of high lake level, especially following inflows from Lalbert Creek. An addition of three metres of water to the lake more than doubles the hydraulic gradient in the Lake Wahpool direction. This semi-regular flushing of the lake into the groundwater system prevents significant halite buildup within the lake basin. It accounts for the lake hydrochemistry, which closely resembles that of the inflowing groundwaters from whence comes the bulk of the salt entering the lake.

That recharge of the Parilla Sand aquifer by the lake occurs under the lake-full regime is confirmed by the relatively low salinity-90,000 to 120,000 mg/l (analysis 9, Table 10.6)-of the underlying reflux brines. Lake salinities of this order only occur during, and for a period after, the wetter seasons, when lake levels are high and hydrostatic heads are especially favourable to outseepage (analyses 1, 2 and 3, Table 10.6).

10.6 The Hydrochemistry of the Lakes of the Raak Boinka -A Comparison

Similar hydrological conditions to those existing in the Tyrrell Basin are found extensively throughout the Sunset Country of far northwestern Victoria (Macumber, 1980). Data from the Raak boinka and Pink Lakes boinka (Figure 6.2) show that the salinities of their respective underlying groundwaters (Table 8.1) are virtually the same as that of the Timboram-Wahpool reflux brines, i.e. about 100,000 to 120,000 mg/l. Like the latter waters, they are essentially concentrated versions of the inflowing regional groundwaters (see analyses 1 and 2, Table 10.7, and the Nangiloc-Colignan waters-Table 7.2).

The Raak and Pink Lakes boinkas differ from lakes Timboram and Tyrrell in that there is no surface drainage system feeding the playa lakes in the discharge complexes. Apart from groundwater inflow, the lakes only receive direct precipitation and some minor runoff from their immediate surroundings. Yet in most respects the systems are similar since in the Tyrrell Basin the feeder creeks rarely flow and for most years their input from upbasin is virtually nil. As at Lake Tyrrell, the groundwater brine and its parent lake water have a predetermined chemical character based on that of the regional groundwater (Tables 10.7 and 10.9). In both the Raak and Pink Lakes boinkas, groundwater emerges in many salinas (Figure 6.9) scattered throughout the boinka depressions (Chapter 6.2), and both halite and gypsum are extensively harvested.

TABLE 10.7 - Hydrochemistry of Raak salinas

Sample	1	2	3	4	5	6	7
TDS (mg/l)	47,266	122,170	205,569	135,723	195,458	249,498	328,137
Chloride	25,140	68,000	111,500	72,200	114,500	147,250	192,000
Bicarbonate	28	37	120	117	31	26	50
Sulphate	4,984	8,987	18,950	13,150	6,466	6,595	10,300
Bromide	79	170	270	130	90	90	175
Calcium	431	550	680	1,170	440	500	820
Magnesium	1,890	3,800	6,560	3,800	2,315	2,315	3,430
Sodium	14,444	40,250	66,800	44,900	71,250	92,500	12,100
Potassium	208	360	680	450	360	218	355
Iron-soluble	57	2	1	1	1	1	2
Silica	1*	10	3	1	3	2	2
pH	5.3	6.2	6.9	6.5	6.1	6.5	6.7
Cl/Br	318	400	413	554	1,270	1,636	1,097
Cl/SO ₄	5.0	7.6	5.9	5.5	18	22	19
Cl/K	121	189	164	160	318	675	541
Cl/Mg	13	18	17	19	49	64	56
Mg/Br	24	22	24	29	26	26	20

* Abnormally low value for SiO₂

1 Inflowing regional groundwater - Mournpoul 10001

2 Raak reflux brine-Bitterang 8001

3 Macarthur's Lake

4 Raak main lake

5 Spectacle Lake

6 Salt Lake

7 Sunray Lake



Figure 10.9 - Spectacle Lake. The lake consists of two interconnected salinas.

The colour difference stems from the presence of *Dunaliella* algae in the right hand salina, while in the left salina iron rich groundwaters from the underlying Parilla Sand aquifer have been pumped from the Mournpoul 10001 bore (analysis 1, Table 10.7) on to the lake floor. The consequent precipitation of iron oxide gives a brown colour to the water (and eventually to the salt being harvested). The Mournpoul 10001 bore, situated on the edge of the lake, has a static water level about 2 m above the lake.

Spectacle Lake lies within a dunefield on the outskirts of the Raak boinka, which can be seen in the background. The main Raak salina can be clearly discerned.

As at Lake Tyrrell, the individual salinas are covered with a shallow sheet of water in winter, but dry out in the warmer months, following the same seasonal cycle of halite re-solution and evaporative concentration whenever a halite crust is present. Each salina may be categorized as either a halite producer with a high Cl/Br ratio when wet (analyses 5-7, Table 10.7), like Lake Tyrrell, or as a through-flushing system, like Lake Timboram, with a lower Cl/ Br ratio (analyses 3 and 4).

It is notable that, while the underlying groundwater brines have salinities of up to 10%, it is common for the very shallow groundwaters immediately beneath the salina floor to be significantly more saline, often being close to halite saturation. Despite this condition, very few salinas are halite producers, which implies that the presence of this shallow interstitial brine pool is not a sufficient condition in itself to produce a halite evaporite body within the overlying lake. A number of additional factors including the hydrostatic closure of the lake to outseepage, are required for a halite body to be formed and then preserved once the lake dries out.

All the analyses, apart from No. 4, come from small ephemeral salt lakes (Figure 10.9), each less than one kilometre in diameter. Analysis 4 is from the main Raak salina at the eastern edge of the boinka; this is an extensive salt lake complex, some tens of kilometres in area (Figure 6.9). While analyses 6 and 7 from Salt Lake and Sunray Lake have the highest salinities, their respective chloride ratios (Table 10.7) show this to be largely due to halite re-solution. It is especially clear in the cases of Spectacle and Salt lakes, where the concentration ratios (analyses 5 and 6, Table 10.7) show the water to be little more than a halite enriched regional water that has undergone only a slight degree of evaporative concentration. By contrast, the less saline water from Macarthur's Lake (analysis 3) shows an average degree of concentration of about 3.5 times that of the regional groundwater. Analyses from Macarthur's Lake (Table 10.8) show that a salinity rise from 149,000 to 206,000 mg/l was entirely due to evaporative concentration.

TABLE 10.8 - Hydrochemistry of Macarthur's Lake

	TDS (mg/l)	Cl ⁻	SO ₄ ²⁻	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	Br ⁻
July	148,785	78,800	14,600	1,000	4,550	49,000	525	180
September	205,569	111,500	18,950	680	6,560	66,800	680	270
Conc. factor	1.4	1.4	1.3	0.7	1.4	1.4	1.3	1.5

Following the final drying of the lake, the residual brines drop below the lake surface to become part of the shallow groundwater system. The extent of the gypsum loss on evaporation is shown by the major depletion of calcium and the lesser depletion of sulphate (on account of its high initial concentration compared with that of calcium).

TABLE 10.9 - Concentration of Raak lakes (lake numbers from Table 10.7)

Lake number	2	3	4	5	6	7
Bromide	2.2	3.4	1.6	1.1	1.1	2.2
Magnesium	2.0	3.5	2.0	1.2	1.2	1.8
Sulphate	1.8	3.8	2.6	1.3	1.3	2.1
Calcium	1.3	1.6	2.7	1.0	1.2	1.9
Potassium	1.5	3.3	2.2	1.7	1.0	1.7
Sodium	2.8	4.6	3.1	4.9	6.4	8.4
Chloride	2.7	4.4	2.8	4.5	5.9	7.6

Table 10.9 shows the concentration of each element in the groundwater and lake waters (numbers 2 to 7 of Table 10.7), with respect to the concentration of each equivalent element occurring in the regional groundwater (analysis 1 of Table 10.7).

It is notable that the Raak reflux brines (analysis 2) have a much lower salinity than the lake waters, with the exception of the main Raak salina. On the basis of the Tyrrell evidence, it is likely that much of the reflux brine was recharged under wetter conditions than exist today, with perhaps any present-day recharge coming largely from the main Raak salina at times of favourable hydrostatic heads. In this respect, the main Raak salina lying on the eastern edge of the Raak Basin may perform a similar role to that of Lake Timboram on the eastern edge of the Tyrrell Basin.

10.7 Examples from the Loddon Valley

A further example of a re-solution lake is Clay Pit Lake, a small salina in the Loddon Valley, about 70 km east of Lake Tyrrell. This lake is on the border between the Mallee and the Riverine Plain. In May 1980, samples were taken from the lake and from a nearby pit sunk into the water table; analyses showed the lake salinity to be 323,500 mg/l and the groundwater to be 76,000 mg/l. However, apart from excess NaCl in the lake, the lake and groundwater chemistries are virtually the same; of note are the Cl/Br ratios of 348 in the groundwater and 1382 in the lake. This lake is a closed system and produces halite.

TABLE 10.10 - Water analyses from Clay Pit Lake—Loddon Valley

	TDS	Cl ⁻	SO ²⁻	Ca ²⁺	Mg ²⁺	K ⁺	Br ⁻	Na ⁺	Cl/Br
Clay Pit Lake	323,542	193,500	5,030	1,350	3,650	63	140	119,700	1,382
Groundwater	76,009	43,500	5,200	1,650	3,800	55	125	21,500	348

10.8 Summary

The geochemical data from northwestern Victoria clearly demonstrates that, where local geography favours groundwater discharge, slight variations in the relationship between water table depth, ground surface level and aquifer salinity strongly influence the interactions between the groundwaters and the lakes. Despite seemingly identical physiographic settings, the effects of these factors on the lakes lead to markedly different hydrochemical characteristics, which may in turn be sufficiently variable to strongly influence the lake biota. While a distinction between resolution lakes and evaporative concentration lakes was first noted in the Tyrrell Basin, it also applies to lake systems throughout northwestern Victoria and can be used for NaCl waters throughout southern Australia.

11. WATER TABLE FLUCTUATIONS BENEATH THE MALLEE SALINAS AND THEIR INFLUENCE ON SALT MIGRATION

11.1 Introduction

During the spring of 1979 diurnal water table fluctuations were noted in a shallow piezometer established on the floor of Macarthur's Lake (Figure 11.1) in the Raak boinka. The piezometer was established to investigate the relationship between the lake water and groundwater during the drying up of the lake. It was inserted into the lake floor to a depth of only one metre and slotted over the interval from 0.3 m below the lake floor to 0.3 m above the lake floor. Water levels were monitored by a Leopold-Stevens continuous water level recorder (Figure 11.2). Almost immediately after the lake had dried out, diurnal water table fluctuations commenced, ranging in amplitude from about 1 cm to 5 cm—the largest fluctuations occurring in mid-summer. The fluctuations continued until the following autumn, when the lake once again held water. Because the fluctuations were virtually from the lake surface, it was considered that the process could play a significant role in the movement of precipitated salts from the lake surface into the shallow groundwater system.



Figure 11.1 – Leopold-Stevens recorder station at Macarthur's Lake in the Raak boinka. At this site the casing is screened for 0.3 m above and below the lake floor so as to monitor either the lake level (when wet) or the water table (when dry).

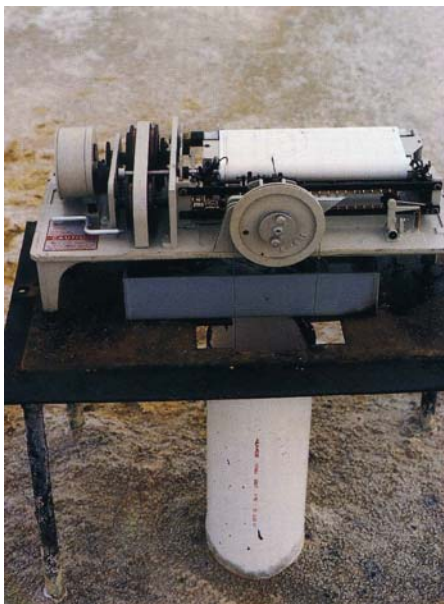


Figure 11.2 – Leopold-Stevens recorder (detail). This recorder is established on the floor of the main Raak salina, about 1 km north of Macarthur's Lake.



Figure 11.3 - Water level recorder situated on the lake floor at southwestern Lake Tyrrell.

The recorder is equipped with a rain gauge. When the water table drops below the lake surface at this site, it has a barometric efficiency approaching 1 - the diurnal fluctuation is superimposed on a six day cycle, brought about by the passage of high and low pressure systems.



Figure 11.4 - Hole dug through the salt crust at Lake Tyrrell. The fluctuating water table, seen here at a depth of about 10 cm, may come to the surface at times of low atmospheric pressure.

In order to determine the extent to which water table fluctuations occurred in other playa lake systems of northwestern Victoria, four recorders were established on lake floor piezometers and were monitored throughout the dry lake phase in 1980-81. In order to gain further data on the fluctuations the recorders were fitted with automatic rain gauges.

Recording stations were established at four places: the lakeward end of the T-P piezometer line in NW Tyrrell; offshore from the salt works at SW Tyrrell (Figure 11.3); on the main Raak salina at the eastern side of the Raak boinka; and at Macarthur's Lake. The Raak sites were chosen to determine whether salina size influences the fluctuations-Macarthur's Lake being only 200 m in diameter, while the main Raak salina is many kilometres long.

Marked lithological differences exist between the two Tyrrell sites: the southern site is underlain by fine - to coarse-grained sand to a depth of 70 m, while the northern site is underlain by clays.

Although not known at the time of establishing the piezometers, the Raak sites are both underlain by a clay sequence, covered by a thin sand sheet about five metres thick in the case of the main Raak salina. Diurnal water table fluctuations were observed at all four sites.

11.2 Character of the Fluctuations

Once the surface water has gone from the lakes the water table falls to about 1 cm below ground surface, at which point small amplitude diurnal fluctuations commence. These become stronger as the daily temperature increases during late spring and early summer. By November 1980, the amplitude of the fluctuations ranged from 1 to 3 cm; throughout the summer period they ranged from 1 to 4 cm and reached as much as 6 cm/day in late February. The diurnal fluctuations are, at times, superimposed on a falling (or rising) water table; however, water levels remained within 20 cm of the surface and usually were within 5 to 10 cm of the surface.

The summer of 1981 was very dry, and the water tables in the Raak boinka gradually fell to about 16 to 20 cm below the floors of the lakes. In early autumn, the strong summer fluctuations began to wane and by mid-March had virtually ceased, with water tables still well below the surface. For instance, at Macarthur's Lake fluctuations continued over the summer period, and the water table gradually fell despite the occasional rainstorms which briefly brought the water back to the lake surface. By early March, the water table was 15–16 cm down, at which depth the fluctuations had virtually ceased. The water table returned to the surface on 31st March after heavy rain, but soon began to fall again, with a distinct 2 cm daily oscillation, until it reached about 11 cm below the lake floor, after which fluctuations ceased.

Unlike the other three sites, the water table fluctuations at the SW Tyrrell site did not cease in March but continued with undiminished amplitude throughout the autumn reaching depths of 25 cm below the surface. As will be shown later, this area has a distinct fluctuation pattern not seen at the other sites.

11.2.1 Diurnal Water Table Fluctuations

In general, water tables fall overnight, reaching their lowest levels in the morning; they rise in mid-afternoon to peak in late afternoon and evening. The oscillations at southwestern Lake Tyrrell may at times lag several hours behind the other sites. The simplest fluctuations for southwestern Lake Tyrrell and the main Raak salina are shown in figure 11.5, covering the period 18–26 February 1981. Water table fluctuations are compared with atmospheric pressure and temperature data from the meteorological station at Mildura. (For comparative purposes the Tyrrell curve, which tends to lag behind the other stations, has been synchronized with the pressure data.) While the fluctuation pattern at Raak has a simple sinusoidal curve, inversely mirroring (note scale inversion) the temperature curve, that at Tyrrell more closely reflects the pressure fluctuation, which is a combination of the temperature induced, diurnal pressure oscillation superimposed on a less regular pattern of high and low pressures. In these examples, the water table fluctuates from 7 to 19 cm below the lake floor at Tyrrell and from 4 to 13 cm at Raak.

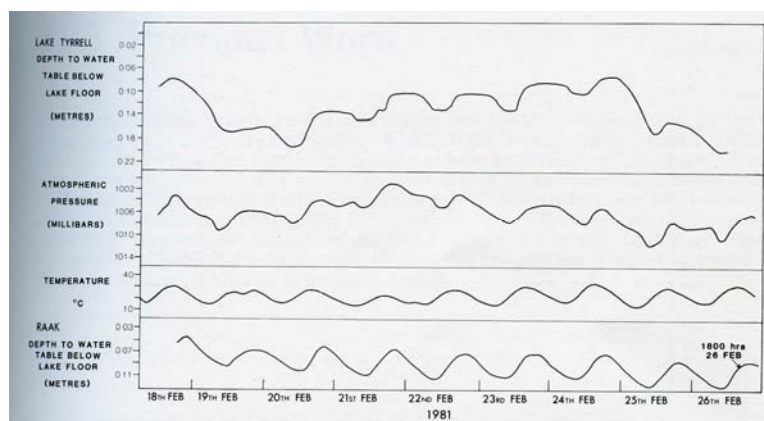


Figure 11.5 - Water table fluctuations at SW Lake Tyrrell and at the Raak main lake, with additional atmospheric pressure and temperature data, February 1981

A further set of data (Figure 11.6) is presented for Macarthur's Lake in the Raak boinka, situated about 100 km northwest of the Tyrrell site. The chart spans the period 21—30 January 1981, when the water table ranged from between 2 and 13 cm below the lake floor, the daily oscillation being from 2 to 5 cm. (The broad peak of 28 January was due to rain.) An unusual feature at Macarthur's Lake during this period is the development of semi-diurnal pressure fluctuations, with two daily maxima and minima. (Similar semi-diurnal peaks can also be faintly discerned from 20 to 22 February on Figure 11.5.) The absence of a comparable set of semi-diurnal water table fluctuations is seen to reflect the role of temperature in the reinforcement of only one of the two sets of pressure oscillations (see Temperature Induced Fluctuations below).

11.2.2 Water Table Fluctuations in Response to Low and High Pressure Systems

An additional feature, most clearly observed in the southwestern Tyrrell site, is the response of the water table to changes in the weather. This is reflected in larger scale fluctuations, often about 8-10 cm but occasionally reaching 20 cm. Under these influences, the record shows a broader six-day cycle, matching the pattern of high and low pressure systems passing across southern Australia. Diurnal water table fluctuations are superimposed on the broader cycles (Figure 11.7). Figure 11.7 shows water table levels at southwest Tyrrell from 6—18 February 1981, compared with air pressure data from the Bureau of Meteorology station at Mildura. The Mildura station lies 150 km to the northwest of Lake Tyrrell, and is therefore ahead of Tyrrell in receiving the weather: this has been adjusted by matching the Mildura data (as a unit) to the Tyrrell fluctuations.

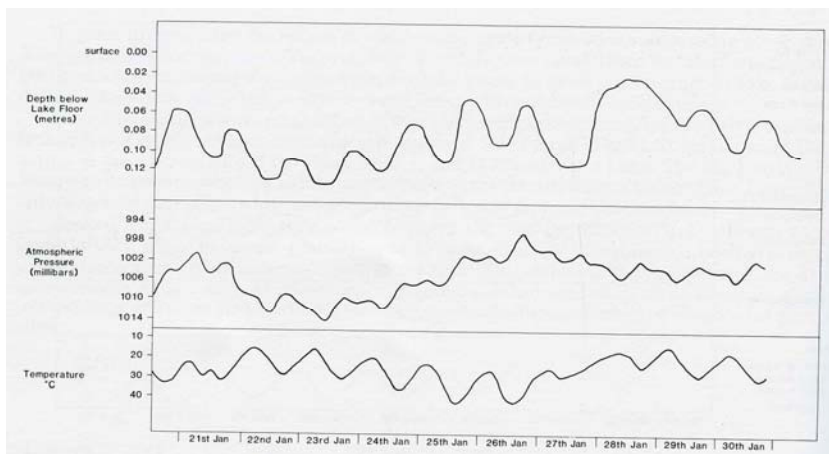


Figure 11.6 - Water table fluctuations at Macarthur's Lake, January 1981

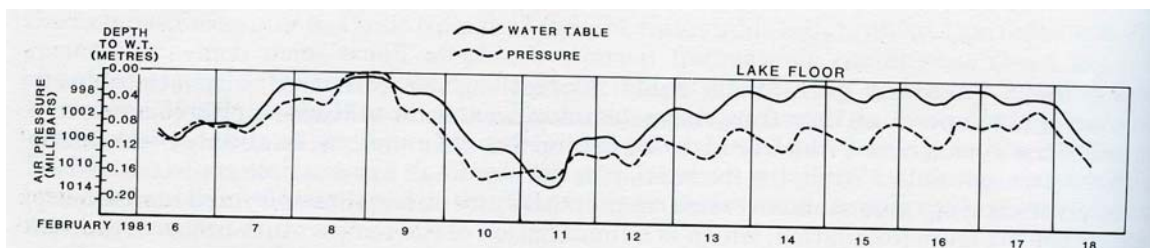


Figure 11.7 - SW Lake Tyrrell water table and barometric fluctuations

Despite the distance between the two stations, there is a remarkably clear match between barometric data and water table fluctuations. Of particular note is the almost 1:1 correlation between barometric pressure and water table fluctuation which is comparable to a barometric efficiency in confined aquifers of about 90-100%. During the 14-day period, the water table fluctuated over a vertical distance of 17.5 cm, being at the surface on the 8-9 and 14-17 February and at a depth of 17.5 cm below the surface on the 11th. Diurnal fluctuations are superimposed on a falling water table on 9-10,th and on a rising water table on the 11-14.th On 14-17th fluctuations with an amplitude of 2 cm from the surface occur. Of prime interest is the fact that the surface during this entire period was formed by a 7 cm halite crust through which the water table fluctuated (Figure 11.4).

11.3 Previous Work

Diurnal fluctuations in both groundwater tables and in stream flows caused by the daily evaporative cycle are well documented (White, 1932; Troxell, 1936; Tromble 1977; and Meyboom, 1967). However, in these instances water tables fall during the day under the high evapotranspiration regime and recover during the evening as evapotranspiration declines. They are at their lowest in late afternoon and highest in the early morning. This is the reverse of the situation occurring in the Mallee lakes and is not a significant process here. Small-scale water table fluctuations, not related to evapotranspiration cycles, have been variously documented. In general, the fluctuations fall into two groups: those induced by temperature changes and those responding to pressure changes.

11.3.1 Temperature Induced Fluctuations

The effects of a temperature gradient on moisture transfer within the soil water zone have been largely the province of soil scientists. Bouyoucos (1915) first showed that soil moisture moves from warmer to cooler soils. This work was confirmed by Lebedeff (1927), Taylor (1962), and Carey and Taylor (1962), who showed water movements from areas of high temperature to areas of low temperature in response to vapour pressure differences, and to matric potential gradients.

Gatewood et al. (1950) noted a sudden fall in water levels, coinciding with a sudden drop in temperatures as a cold front passed over tanks where experiments into water usage were being carried out. They showed that the addition of heated stones to the soil surface caused a sharp rise in the water table without any significant barometric change. Tison (1954) considered that the water level fluctuations observed by Gatewood et al. resulted from changes in surface tension on heating, causing the migration of capillary water into the water table, which consequently rose.

Meyer (1960) noted that as early as 1915 he had observed temperature induced fluctuations in water tables while he was experimenting on capillary rise in fine-grained sediments. Water levels rose when the soil column was warmed and fell when it was cooled. He noted similar effects in the field, in Minnesota, where winter lowering of temperatures was accompanied by a fall in the water tables, while spring warming saw a rise in water tables. He also observed diurnal fluctuations in shallow wells during winter: low levels being at 9.00 a.m. and high levels at 4.00 p.m., the extent of the fluctuation being roughly proportional to the daily temperature range. Meyer (1960) comments that the amount of water held by capillarity is greater in cold soils than in warm soils. He explains the fluctuations by the transfer of moisture between the capillary zone and the water table in response to temperature changes. Meyer considers the process to be primarily a migration of liquid water, induced by changes in soil temperature. The rapidity and extent of water transfer, he contends, could not be achieved by processes of vapour diffusion. This conclusion was also reached by Jumikis (1962). Diurnal fluctuations in shallow water tables under the Great Salt Plains of Oklahoma were observed by Davis (1968). In June 1968 the daily fluctuation averaged 7–10 cm, while the maximum depth to the water table was only 30 cm; daily maxima occurred between 6.00 p.m. and 9.00 p.m. while minima were between 8.00 a.m. and 10.00 a.m. Davis correlated the water level fluctuations with diurnal changes in temperature and relative humidity.

11.3.2 Atmospheric Pressure Effects on Shallow Water Tables

The effects of atmospheric pressure changes on water levels in unconfined aquifers are not as clearly defined as the effects on confined aquifers. However, Tuinzaad (1954) shows that where the capillary fringe lies within reach of the surface, atmospheric pressures can influence water tables. Peck (1960) showed that water level changes may be caused by the effects of pressure changes on air entrapped within the shallow groundwater. This study was confirmed experimentally by Norum and Luthin (1968), who considered that the pressure induced water table fluctuations were larger than those predicted by Peck's theoretical study. Bianchi and Haskell (1966) noted water table fluctuations when carrying out recharge experiments. They considered these to be caused by barometric pressure on air entrapped within the aquifer. Van

Hylckama (1968) observed diurnal fluctuations in evaporimeters, which he attributed to temperature induced pressure fluctuations. He considered, however, that half of the fluid occupying the pore space in the evaporimeter must be bubbles in order to provide sufficient air space to accommodate the observed barometric rise and consequent water table fall. An entirely different mechanism was suggested by Gilliland (1969). He proposed a rigid plate model as an alternative explanation to that of Peck for fluctuations in water table wells. He shows that if sufficient water is present, the soil water zone may act as a rigid plate for atmospheric pressure fluctuations. The water is suspended between the grains by surface tension. If a continuous layer exists, "any change in atmospheric pressure will be transmitted directly to the aquifer structure by the increased surface tension forces". If it is assumed "that the adjustment to pressure changes is instantaneous, the water table will rise or fall instantaneously". The existence of a rigid plate requires that the grain size of the surficial material satisfies the equation:

$$p < 4T/\partial Pa \quad (11.1)$$

Where T is the surface tension of the water
 p is the effective pore opening and is approximately equal to the grain size
 ∂Pa is the change in atmospheric pressure

Gilliland considers that the increase in barometric efficiency of water table wells, caused by the entrapment of air in the soil zone by infiltrating precipitation (the Lisse effect of Meyboom, 1967), is a manifestation of the rigid plate model. He concludes that the barometric efficiency of a well is related not only to the properties of the aquifer but also to the properties of the overlying rigid plate and to infiltrated precipitation - the soil moisture zone is one manifestation of the rigid plate. An attempt to accommodate all the observations was made by Turk (1975), who had previously recorded similar diurnal fluctuations from beneath the Bonneville Salt Flats. Water tables fluctuated from 1.5 to 6 cm daily in summer and from 0.5 to 1.0 cm in mid-winter. Highest water tables were in late afternoon, with lowest in mid morning. Turk attributes the fluctuations to temperature induced atmospheric changes acting upon the capillary zone.

11.4 Discussion

In summary, three types of fluctuation are described above:

- a. The diurnal fluctuation, with water tables high in late afternoon and low in the early morning (Meyer, 1960; Davis, 1968; Van Hylckama, 1968; Turk, 1975);
- b. Falls in water tables related to seasonal cooling (Meyer, 1960; Turk, 1975);
- c. Sudden falls in water tables in response to the passage of cold fronts (Gatewood et al., 1950).

The hydrology of the Mallee salinas differs from the above described sites in that water tables are only occasionally more than 20 cm below the surface and commonly oscillate within 2-4 cm of the surface. In the latter instances, the thickness of the capillary zone is less than 4 cm. In winter the salinas have a shallow water cover. A feature of the oscillation is the tendency for to attenuate at both the tops and bottoms, giving a considerable period in which water tables are static prior to reversing their direction. However, with the exception of the SW Tyrrell recorder the oscillations closely conform to the pattern of diurnal fluctuations described elsewhere. The feature not previously described in the literature is the larger amplitude oscillation linked to the passage of high and low pressure systems-this was, however, theoretically predicted by Peck (see below).

11.5 Cause of Fluctuations

Bouwer (1978) noted that water levels in piezometers do not always record the true level of the water table. This situation arises when there is a restricted transmission of atmospheric pressure change through the vadose zone to the water table, while at the same time a piezometer open to the atmosphere directly responds to the change. At the SW Tyrrell site, the near 1:1 correspondence between the barometric pressure and the water table fluctuations is in close accord with the theoretically derived values of 13.6 cm fall in water level with each 13.3 millibar rise in atmospheric pressure. The piezometer is clearly acting as a barometer and has a barometric efficiency which comes close to 100%. Since the water table is always within 25 cm of the surface, a simple test was applied to determine whether the piezometer reflects processes operating merely within the piezometer pipe or whether it shows a more general water table fluctuation. When the water table is shown as being at the surface (and at such times the salt crust is notably wet), a number of holes dug 20 cm into the lake floor filled with water, i.e. the water table is at the surface. Conversely, when the water table is shown as being well below the surface, holes dug through the salt crust show it to be in the sediment below the salt crust. They do not fill but instead show the same fluctuation as in the case of the water level recorder.

As Turk (1975) points out, the fluctuations may result from both temperature induced surface tension changes and temperature induced pressure changes - the processes are reinforcing and not separable in the field. They would quite adequately account for all previously observed diurnal fluctuations; however, in the case of the Mallee discharge lakes, Turk's mechanism of water movement between the water table and the capillary zone has some unusual implications. These stem from the limited extent of the capillary zone when the water table is only about 1-2 cm below the lake floor at the time when fluctuations first appear. This is about the same magnitude as the daily fluctuation, therefore, any explanation based on the transfer of water between the water table and the capillary zone requires that the capillary zone must accommodate a further 2 cm of water as the levels fall. This implies that the capillary zone, occupying an interval only a few centimetres above the water table, is at best only half saturated prior to the daily fall and absorbs a further two or more centimetres of water as the water level drops.

The process is even more unlikely when, as happened at southwest Lake Tyrrell in February 1981 (Figure 11.7), the oscillations were from the surface, and no capillary zone existed prior to the fall in water tables. Under these conditions, the alternative explanation of temperature induced pressure changes acting on air entrapped within the phreatic zone (Peck, 1960) becomes a more attractive mechanism. Indeed, a number of the characteristics of the water table fluctuations under the ephemeral Mallee salt lakes closely fit the theoretical and experimentally based concepts expounded by Peck. For instance, the almost 1:1 correspondence between water table and atmospheric pressure oscillations observed at Lake Tyrrell following the passage of high and low pressure systems comes close to the predicted values of $\partial Z/\partial P = -1/\rho$ when the water table is at the surface. Peck comments: "As the passage of weather disturbances (highs and lows) is associated with air pressure changes of as much as 50 gm wt cm⁻² in not uncommon cases, considerable variation in the position of a water table near the surface can be expected". Additional details of Peck's work are given hereunder, with a short discussion on its implications when applied to ephemeral playa lakes. Peck provided a theoretical analysis for determination of the magnitude of water table fluctuations in response to atmospheric pressure acting on air entrapped in the water. The theoretical predictions were strongly supported by experimental data. Most relevant to this study is the observation that the maximum rate of change of water table height with air pressure occurs when the water table is at the surface - this situation arises once the surface water has gone from an ephemeral playa lake.

Commencing with the gas equation for entrapped air (applicable where the water table is within 900 cm of the surface),

$$V = CT/(P + \rho (Z-z))$$

(11.2)

Peck (1960) derives an equation for the water table fluctuation in response to atmospheric pressure changes.

$$\partial Z / \partial P = -CTh / \{(\sigma - \Phi_h)(P + \rho Z) + \rho(P + \rho(Z-h)) + \rho CTh\}$$

(11.3)

Where C is the quantity of air entrapped in a unit volume of sediment

h is the height of sediment column measured from an impervious base

V is the volume of air entrapped in a unit volume of soil

P is the external air pressure

T is temperature

Z is the height of the water table above the impervious base

z is a vertical ordinate, positive upwards, with its origin at the base of the sediment column

σ is the porosity of the sediment

Φ is the apparent volumetric water content at 'h' (= the volumetric water content plus the entrapped air)

ρ is the density of water in the sediment.

11.6 Implications of Peck's Equation for Water Table Fluctuations

Peck's equation indicates that a maximum response of the water table to atmospheric pressure change is:

$$\partial Z/\partial P = -1/\rho, \text{ when } (\sigma - \Phi_h)(P + \rho Z) P + \rho(Z-h)] \text{ becomes zero.} \quad (11.4)$$

This is obtained when Z equals h , i.e. the water table is at the surface, and Φ_h equals σ , i.e. the volumetric water content including entrapped air equals the porosity.

Peck confirmed experimentally that $\partial Z/\partial P \rightarrow -1/\rho$ as $Z \rightarrow h$. More recently, McWhorter and Sundae (1977) repeated Peck's analysis, adding that thick aquifers should exhibit larger water table fluctuations compared to those occurring in thin aquifers. Much greater aquifer thickness, more than any other factor, distinguishes the site at southwest Tyrrell from the other sites, where at best only a thin sand sheet covers an essentially clay sequence.

Thus, at southwest Lake Tyrrell, near optimum conditions for pressure induced water table oscillations occur once the surface water vanishes from the lake floor. Given a range of groundwater densities from about 1.1 to 1.2 g/cm³, the barometric efficiency while the water table remains close to the surface would be approximately 85 to 90%. This is certainly the case at southwest Lake Tyrrell. However, in other situations, e.g. Macarthur's Lake in December 1980, the passage of lows and highs is not so much reflected by a sympathetic overall rise and fall in water tables, but instead by an increase and decrease in the amplitude of the diurnal oscillation, with the water table returning to within a few centimetres of the surface each day. The fluctuation pattern becomes more complicated with the addition of rain to the lake floor. Although the value of $\partial Z/\partial P$ changes very slowly as the water table falls below ground surface, a critical depth is reached whereupon there is a marked falloff in the amplitude of the water table fluctuation in response to a rapid change in ($\sigma - \Phi_h$). In Peck's example this occurs at a depth of 25 cm, however the decline in $\partial Z/\partial P$ soon levels off, to be virtually independent of water table depth—at about 35 cm below the surface.

This process may explain the virtual lack of significant water table oscillations in the recorders situated in the Raak boinka during the autumn of 1981, when water tables had fallen to several tens of centimetres below the lake floor. This seems to be a depth dependent feature, given the re-imposition of an oscillatory pattern once water levels again approach the surface following rain storms. While the southwestern Lake Tyrrell recorder faithfully conforms to the Peck model, the other recorders show characteristics which conform to this model to varying degrees at different times. At other times they depart from the model and may then be examined in terms of other models already discussed. This variation is not unexpected, given the differences in lithology and hydrology of the different sites and the roles of atmospheric pressure, temperature and precipitation.

11.7 Summary

In summary, therefore, all recorders show water table fluctuations beneath the dry playa lakes, the fluctuations commencing almost immediately after the lake dries out. The fluctuation patterns vary considerably—both between recorders and temporally at each recorder site. Since a single recorder gives a whole range of responses under differing combinations of conditions, there is no lack of data linking the fluctuation pattern to various factors such as atmospheric pressure, temperature, precipitation and water table depth. The problem where no single model fits is to disentangle the dominant factor or combination of factors operating at any one time. However, while the interpretation of data from the playa lakes should provide a further understanding of the interrelationships existing between the various factors, this must await further investigation.

11.8 Water Table Reflux the Influence of Water Table Fluctuations on the Reflux of Evaporite Salts

Undoubtedly, the most significant implication of the water table fluctuations for this study is the transference of salts from the lake surface into the shallow groundwater systems. In the case of the halite saturated re-solution lakes, halite precipitation commences almost immediately that evaporitic concentration gets underway: at Lake Tyrrell this builds up a 7 cm thick halite crust by the time the lake finally dries up. As the lake shrinks, the surface level of the brine gradually falls until it drops below the lake surface; it then becomes the water table at the top of the phreatic zone. At this point, diurnal fluctuations commence, with the water table 1 or 2 cm below the lake floor. Where a salt crust is present, the top of the crust acts as the lake floor, and the water table fluctuates through the newly deposited halite (Figure 11.4).

The lake waters do not normally reach the stage where magnesium salts precipitate before the water level drops below the lake floor effectively preventing further evaporative concentration. Should magnesium (and potassium) salts be precipitated during a snap drying of the lake under extreme temperature conditions, these salts would be quickly redissolved from the salt crust by the fluctuating groundwaters and thus lost to the surface. Chemical analyses of late-stage lake waters and the resulting salt crust and shallow groundwater brines are given in Tables 10.3, 10.4 and 10.5.

Within the throughflow lakes, halite precipitation only begins at a late stage of evaporation after halite saturation is reached; this may be preceded by a gypsum phase, during which gypsum is precipitated as a sugary textured mush (cf. Teller et al., 1982). At best, only a very thin salt crust remains when the lake dries out. In these instances, the shallow groundwaters are rarely saturated with NaCl and each time the water table reaches the surface, halite is removed. This process is accelerated by occasional rainstorms, which, apart from their leaching qualities, cause the water table to rise rapidly to the surface where it may remain for several days before again subsiding. Ultimately, the only salt left at the surface is a thin efflorescence due to capillarity.

Thus, the reflux of the salts from the lake surface into the shallow aquifer is assisted by diurnal water table fluctuations, which may in some instances be superimposed on a broader (approximately six-day) oscillation having a larger amplitude. The reflux process is essentially a hydrodynamic one, acting only after the lake has dried out. Where a permanent salt crust exists the oscillation of the water table through the crust removes any salts with which the shallow groundwaters are not already saturated. This process explains the absence of any significant deposits of magnesium and potassium salts from the thin evaporite bodies formed in playa lakes of the Murray Basin (see Chapter 10.3).

The process of summer reflux of evaporite salts by an oscillating water table under lake dry conditions (*water table reflux*) is markedly different to winter or lake-full reflux, when hydrostatic head provides the energy for outseepage of both salts and lake water.

12. GROUNDWATER HYDROLOGY OF THE TYRRELL BASIN

12.1 General-Salinities and Surface Levels

Groundwater occurring in the regional Parilla Sand aquifer across much of northwestern Victoria has a fairly uniform salinity, averaging about 35,000 mg/l to 40,000 mg/l, with a density of about 1.03 g/cm³. Water of this salinity is regarded as 'salt water' in most groundwater flow systems. It is certainly the case in coastal aquifers where almost all recorded investigations of salt water intrusion have been carried out. However, in the vicinity of salt lakes the salinity of groundwater is commonly greater than 100,000 mg/l and may reach as high as 290,000 mg/l, as found within the Lake Tyrrell sub-basin. By comparison with these groundwater brines, the regional groundwaters may be considered 'fresh'.

The salinity of the regional groundwaters is relatively uniform throughout the area, and it is on these regional waters that the various groundwater measurements such as static levels are taken to provide data for the construction of water table and potentiometric surface maps for flow system analysis. Therefore, when considering an approach to the problem of ground-waters with varying densities in northwestern Victoria, it was decided that it was more practical to relate all density measurements to that of the existing regional groundwater (S.G. 1.03), rather than putting the whole system on a theoretical basis by using the alternative fresh water value of S.G. 1.00. Therefore, in this study, a density of 1.03 g/cm³ is taken as the 'fresh water' density and all calculations are made on this basis. In the discussions on saltwater and freshwater interfaces, 'salt water' refers to the undiluted reflux brines, 'fresh water' to the regional groundwaters (S.G. 1.03), and a third term, 'transitional water' (i.e. diffuse water) is the water in the transition zone between the fresh and salt waters.

Throughout the text, elevations are given in terms of the Australian Height Datum (AHD), which refers to a standardized height above the Australian mean sea level. In certain instances, for the purposes of clarity, levels are also given relative to the floor of Lake Tyrrell (taken as 41.3 m AHD), and occasionally relative to the floors of lakes Wahpool (45.0 m AHD) and Timboram (46.5 m AHD).

Within the Tyrrell Basin, the overall direction of the regional groundwater flow in the Parilla Sand aquifer is from east to west (Figure 12.1). The water table elevation is 51 m (AHD) in the SR 89 piezometer, situated about 30 km east of the eastern shoreline of Lake Tyrrell, falling to 42 m some 16 km west of the shoreline. This gives a hydraulic gradient of about 2×10^{-4} across the basin. Water table contours (Figure 12.1) in the Tyrrell Basin are strongly affected by the three discharge lakes—Tyrrell, Timboram and Wahpool. The floors of all three lakes intersect the water table, and groundwater discharges from the Parilla Sand aquifer into the respective lake basins. The closed contours around Lake Tyrrell and Lake Wahpool indicate that groundwater inflow is from all sides (Figure 12.6); Lake Timboram, however, does not have a similar closure, and no significant inflow enters from the west.

12.2 Groundwater Divides in Temperate and SemiArid Regions

The closure of the water table contours establishes low groundwater divides on Lake Wahpool and on the downbasin (western) side of Lake Tyrrell (about 3.7 m high), but no comparable divide exists on Lake Timboram.

The presence (or absence) of a divide has considerable implications for groundwater/lake interactions, since it greatly influences the extent (if any) of leakage from the lake into the underlying aquifer. This was shown by Winter (1976, 1978), who in a numerical simulation study of steady state groundwater flow (with uniform density) near lakes demonstrated that the continuity of the boundary of the local flow system beneath a lake is the factor that controls the interaction of lakes and groundwater. If continuity exists, no leakage occurs from the lake (lakes 1 and 2, Figure 12.2). While the altitude of the water table divide on the downbasin side of the lake is one of the most important controls over the continuity of the local flow system, its presence alone does not necessarily ensure continuity. Other influential factors include the position and shape of localized aquifers beneath the lake, the ratios of vertical to horizontal hydraulic conductivities beneath the lake, and the depth of the lake.

Where seepage from the lake occurs, its extent depends upon a combination of the above factors. For instance, where no water table divide exists on the downbasin side of the lake, throughflow conditions prevail. In semiarid regions seepage from a terminal lake helps prevent extreme salinity buildup in lake waters and on the lake floor when the lake dries out. This process is enhanced by surface inflow, which raises lake levels and hence hydrostatic heads, thus aiding leakage. In lakes normally closed to seepage losses, salinity steadily increases, resulting in brine lakes which, as a result of seasonal halite re-solution, remain close to the point of halite saturation for most of the year. This is much less likely to be the case in through-flushing lakes that reach saturation only in the summer months, if at all.

In some instances, especially in the case of small lakes, the hydrological balance may be extremely delicate such that seasonal water table fluctuations may be sufficient to cause throughflow characteristics at one time of the year and produce a closed lake at another (cf. Meyboom, 1967). The same result can be obtained by seasonal lake level fluctuations.

A distinction can be made between the situation described by Winter and that found in the semiarid regions of northwestern Victoria. While the former study investigates lake-groundwater interactions, mainly in temperate areas of moderate relief, the situation existing in this study is one of low relief, very low hydraulic gradients and sparse rainfall. Instead of having a well-developed local groundwater flow system as is the case with the simulation studies, in the semiarid areas of this study the contribution of local recharge to the groundwater divide is minimal, with most of the water being derived from upbasin via the regional groundwater flow system. The small amount of recharge that does occur is spread evenly over the area between the groundwater divide and the lake edge, and in the case of Lake Tyrrell there is no surface water divide that relates in any way to the groundwater divide.

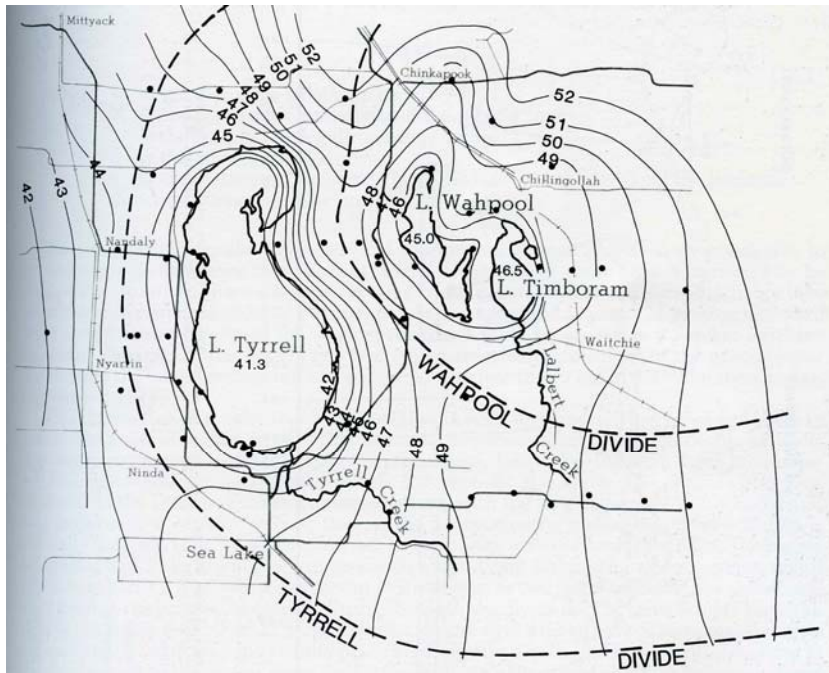


Figure 12.1 - Potentiometric surface of the uppermost parts of the Parilla Sand aquifer in the Tyrrell Basin, with additional potentiometric data from the Blanchetown Clay (based partially on data supplied by Rural Water Commission)

The creation of a groundwater divide in semiarid or arid regions is analogous to the situation where a bore is pumped on a sloping water table. The flow net (Figure 12.3) shows a typical groundwater divide and a stagnation point. The divide is a streamline, and the entire groundwater discharge from the bore is derived from the area contained within this stream-line. In the case of the Mallee lakes, however, the place of a single bore is taken by a groundwater discharge lake, which under the high evaporative regime acts as a natural groundwater pump or series of pumps. In the case of a borehole, the distance to the groundwater divide is given by the following equation (Todd, 1980; Bear, 1979):

$$x_0 = -Q / (2 \pi K i b) \quad (12.1)$$

where

Q	is the discharge from bore (m^3/d)
K	is the hydraulic conductivity of the aquifer (m/d)
i	is the regional hydraulic gradient (m/m)
B	is the aquifer thickness (m)
X	is the distance of the divide from the bore downbasin along the principal direction of regional flow (m)

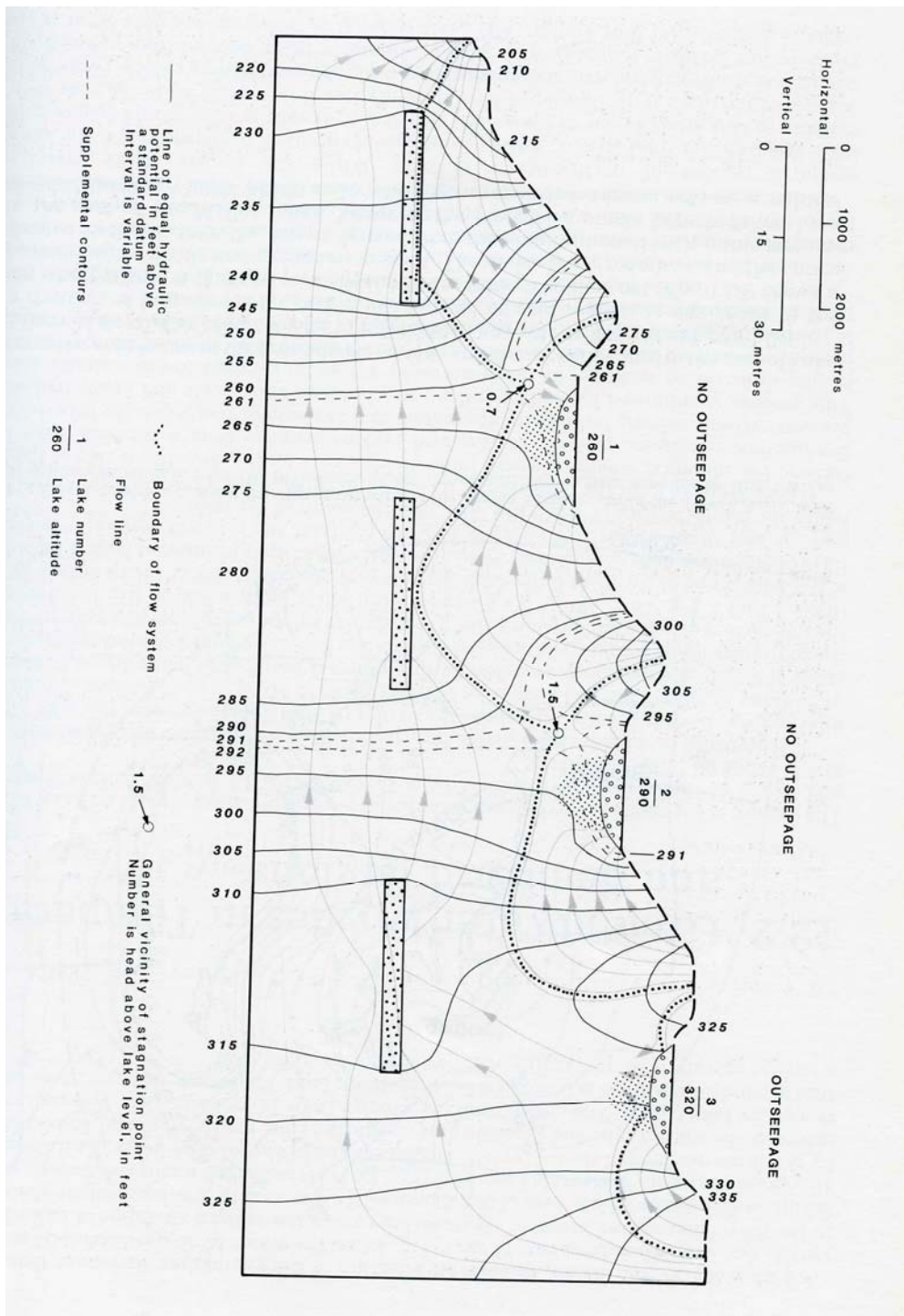


Figure 12.2 – Hydrologic section showing a quasi-quantitative flow net of groundwater flow near lakes in a multiple lake system that contains aquifers (from Winter, 1978)

Low regional hydraulic gradients and low hydraulic conductivities are conducive to groundwater divides being more distant from a borehole. The same generalizations may be made for a situation where a bore is replaced by an ephemeral groundwater discharge lake or salina in an

evaporitic regime. For instance, in the case of Lake Tyrrell, on passing lakeward from the groundwater divide the slope of the water table is only 0.2 to 0.3 m/km until the immediate vicinity of the lake is reached. The corresponding distance of the groundwater divide from the lake shoreline is 6 km, compared with less than 1 km in the simulation models (cf. Winter, 1978).

One distinction that exists between the lakes in semi-arid areas of low relief and those in temperate areas of higher relief is that in the former areas there is no significant vertically downward component to the flow system in the region close to the divide; instead, groundwater flow is essentially horizontal towards the lake. At the divide the groundwater flow conforms to the Dupuit–Forcheimer assumptions, with the flow being essentially horizontal and uniform in a vertical section; that is, the potentiometric head along a vertical line is virtually constant (Figure 12.4). By contrast, in temperate areas of moderate relief, the higher horizontal hydraulic gradients, together with a significant local recharge component, result in a strong vertical downward component at the divide. In this latter instance, the potentiometric head at the base of the aquifer directly beneath the divide maybe considerably less than that at the water table (Figure 12.5). By contrast, in the semi-arid situation, equipotential lines are essentially vertical in the vicinity of the groundwater divide, and the potential existing at the divide must be the closing potential for the basin. It is the potential which density-induced hydrostatic head, produced within the aquifer beneath the lake, must overcome before outflow gradients are achieved.

12.3 Regional Groundwater Flow into the Tyrrell Basin

12.3.1 Flow Net Analysis and Salt Budget-General

The pattern of groundwater flow in the Tyrrell Basin shows considerable variation in the spacing of the water table contours with the steepest gradients occurring near the perimeter of the lakes, where the water table slopes sharply toward the incised lake floor (Figure 12.1). Variations in hydraulic gradient away from the vicinity of the lakes reflect heterogeneities in both thickness and transmissivity within the Parilla Sand aquifer. In the northern and central areas of the Tyrrell Basin the thickness of the Parilla Sand is reduced by the presence of the overlying Blanchetown Clay, which is about 20 m thick in the north but wedges out in the south. The Blanchetown Clay partially cuts into the Parilla Sand with a consequential reduction in aquifer thickness by up to 30% (as shown by a comparison of the Pier Millan 10001 bore situated at the northwest end of the lake and the Bourka 1 bore at the southwest end).

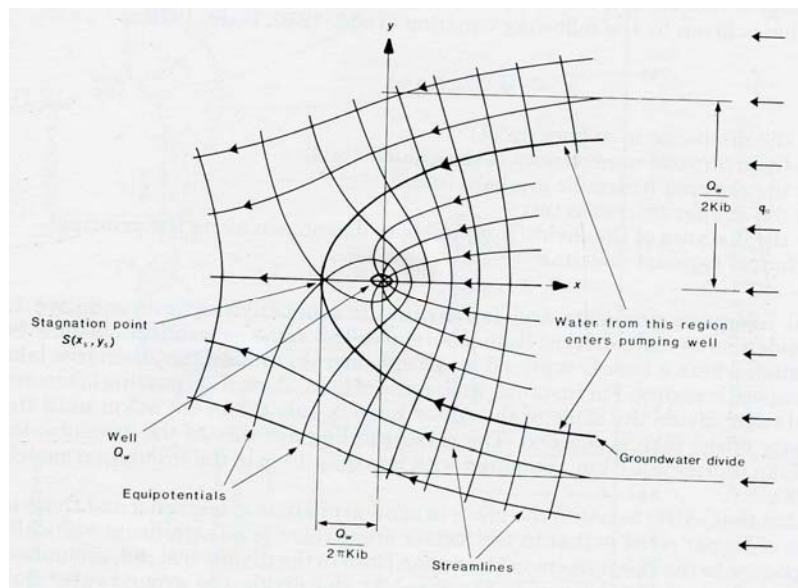


Figure 12.3 - Flow net around a single bore pumping on a sloping water table in uniform flow (after Bear, 1979)

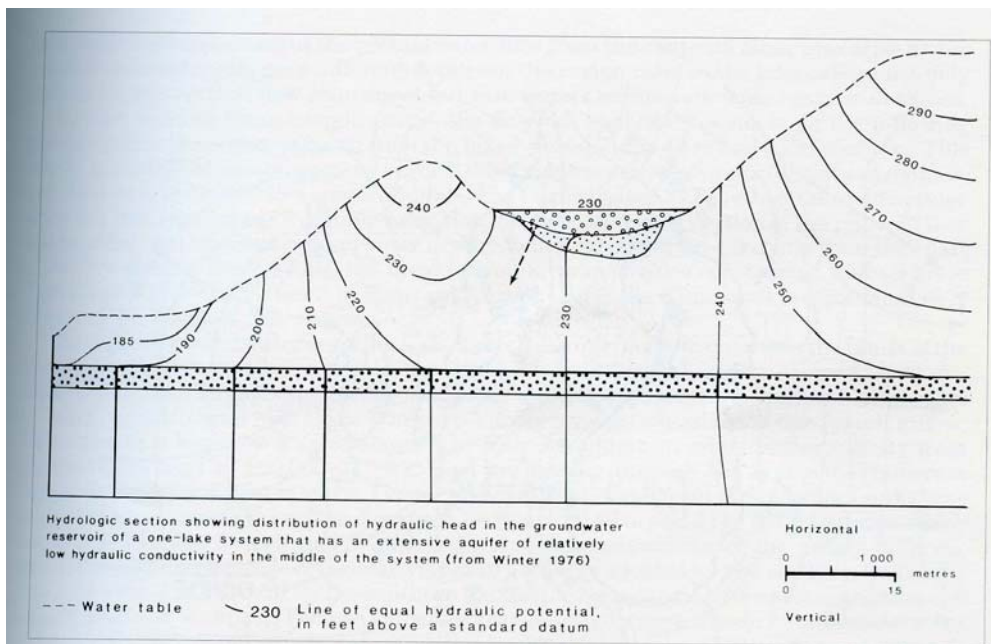
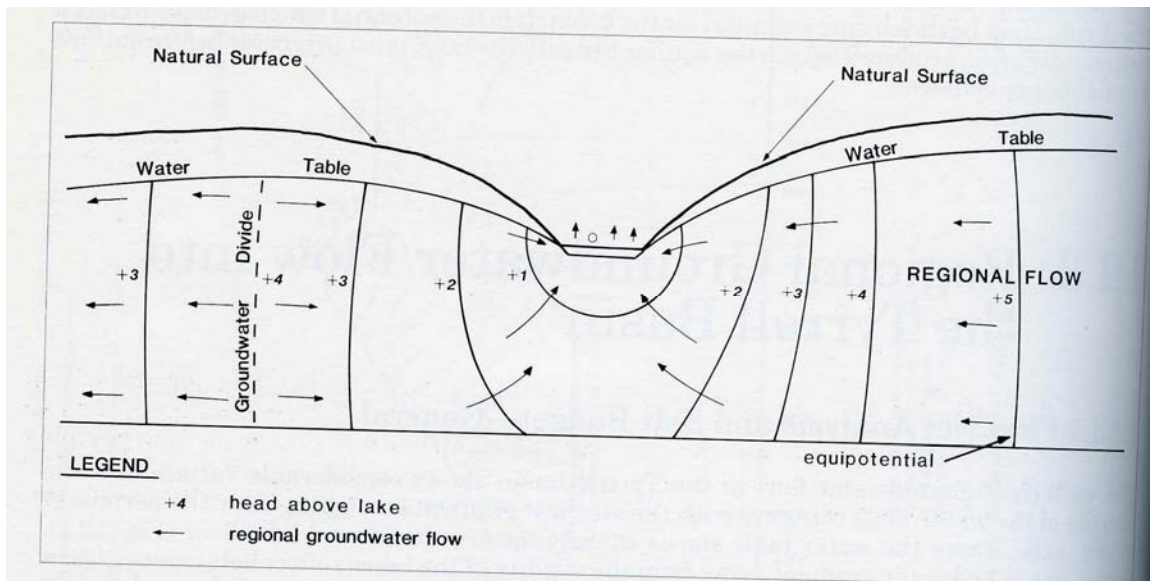


Figure 12.5 - An example from Winter (1976) of the distribution of hydraulic head around a lake open to outseepage.

Unlike the 'semiarid' or 'arid' situation given in Figure 12.4, the hydraulic potential on the divide is significantly higher at the water table than at the aquifer base. In this example, outseepage occurs despite the presence of the groundwater divide.

	Pier Millan 10001 depth (m)	Bourka 1 depth (m)
Blanchetown Clay	0-23	absent
Parilla Sand	23-68	2-67
Geera Clay	68+	67-242
Parilla thickness (m)	45	65

Variations in hydraulic conductivity reflect the broad distribution of the grain size within the Parilla Sand, which ranges from gravel to fine sand and silt. The combination of the two effects is seen in the unusually wide range of values for transmissivity obtained from pumping tests in the Tyrrell district, with values ranging from an abnormally high value of about 700 m²/d at the southern end of Lake Wahpool to about 50 m²/d in the Pier Millan 10001 bore (Table 12.1).

TABLE 12.1 - Transmissivity of the Parilla Sand aquifer

Position	Transmissivity (m ² /d)
NW Lake Tyrrell-Bimbourie 4/5 bores	81 (average of four values)
NW Lake Tyrrell-site 3	200 (Cheetham Salt, unpub.)
NW Lake Tyrrell-Pier Milan 10001 bore	52 (ICI, unpub.)
Tyrrell Creek, SE of Lake Tyrrell 1 bore	170
SW Lake Tyrrell -Bourka 3 bore	168 (average of two values)
From Lawrence (1975), based on a saturated 63 (average of six sets of data) aquifer thickness of 50m	
SW Lake Wahpool	700 (ICI, unpub.)

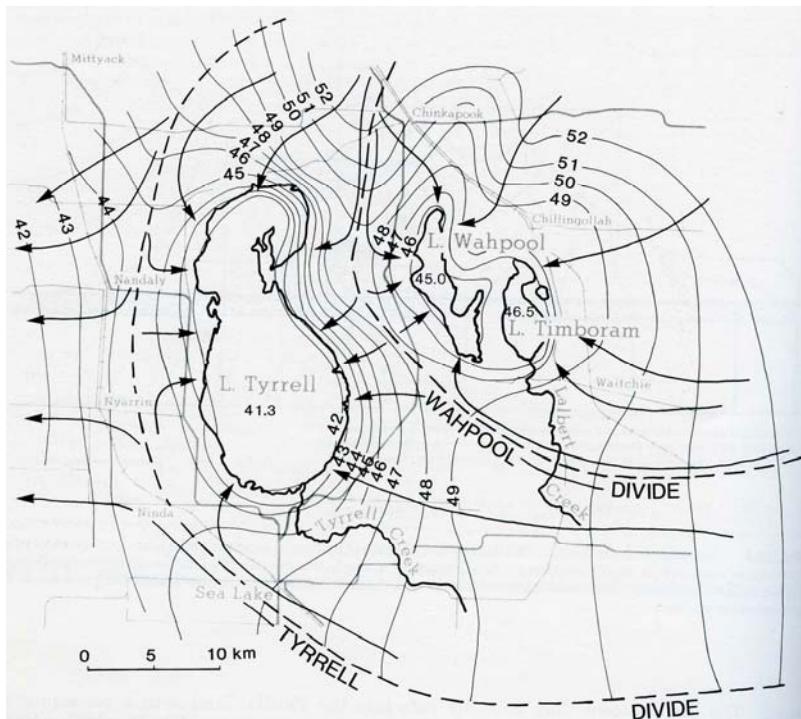


Figure 12.6 - Groundwater flow regimes in the Tyrrell Basin

Since the contours shown on Figure 12.1 represent data from different aquifers, in many instances from piezometers constructed across intervals where water was first encountered, the figure is essentially a map of the water table in the Tyrrell Basin. Difficulties in the quantitative interpretation of the groundwater flow from the contours alone also arise as the result of

piezometers tapping different depths in the region close to the lake, where not only is there a major vertical flow component but also waters with a very wide range of densities.

In order to avoid these complications, the flow net analysis was made on the inflowing groundwater in the region upbasin from the lakes, between the 49 m and 51 m contours. This has the advantage of minimizing any vertical flow component. It eliminates density variations, and the flow is restricted to a single aquifer—the Parilla Sand. The method takes advantage of the fact that each lake is a groundwater sink, fed by a restricted portion of the regional flow and bounded by a localized groundwater divide. Water can enter the lake only from that part of the flow domain lying within the divide. Groundwater divides constructed around lakes Tyrrell and Wahpool are shown in Figures 12.1 and 12.6 (Lake Timboram is considered here as part of the Wahpool flow regime).

The regional groundwater entering Lake Tyrrell can only come from within the limits of the Tyrrell divide, less that lying within the Wahpool divide and therefore captured by Lake Wahpool and Lake Timboram. Since groundwater underflow passes from Wahpool–Timboram to Tyrrell, an additional 33% of the Wahpool–Timboram inflow is added to the Tyrrell inflow. This estimate is based on a salt balance which shows a jump in groundwater salinity from 30,000–40,000 mg/l to 100,000–120,000 mg/l on passing through the Wahpool–Timboram region. Groundwater flow into the Tyrrell and Wahpool–Timboram groundwater sinks was calculated using those areas of the flow path between the 49 m and 51 m AHD contours which are bounded by the Tyrrell and Wahpool divides. In consideration of the widely differing transmissivities, an average transmissivity of 70 m²/d was adopted for the northern half of the area, and a value of 170 m²/d for the southern parts—the boundary is taken as passing through lakes Tyrrell and Wahpool. The two values for transmissivity were chosen after consideration of the available pumping test data and the basin stratigraphy. Although some error is introduced by taking average values for transmissivity, this does not significantly modify any conclusions based on these chosen values.

12.3.2 Flow Net Analysis

Methods of flow net analysis are described in various hydrological texts, e.g. Todd (1980), Freeze and Cherry (1979), and Bear (1979). A flow net analysis provides a graphical solution of a flow pattern composed of two families of curves: the equipotentials (representing contours of a potentiometric surface or a water table) and the streamlines or flowlines (representing the path taken by groundwater as it moves through the aquifer in the direction of decreased potentiometric head). The flow net is constructed so that a set of curvilinear squares is formed from the two sets of lines.

The flow through a net is given by the equation (Bennett, 1962):

$$Q = -T (nf/nd)h \quad (12.2)$$

where Q	is the flow through the net (m ³ /day)
T	is the transmissivity of the medium (m ² /d)
nf	is the number of flow channels
nd	is the number of potential drops
h	is the total potential drop across the net (m)

A. Flow to the Timboram–Wahpool depressions

From Figure 12.7

nf = 21—made up of 12 tubes where T is 70 m²/d,
and 9 tubes where T is 170 m²/d;
nd = 2, and h = 2 m.

$$\begin{aligned}
 Q &= -T (nf/nd)h \\
 Q &= 2,400 \text{ m}^3/\text{d}
 \end{aligned}
 \tag{12.3}$$

B. Flow into Lake Tyrrell

Flow coming from south of the Wahpool divide (Q_s): $nf = 2$, and $T = 170 \text{ m}^2/\text{d}$.

Therefore, $Q_s = 340 \text{ m}^3/\text{d}$.

Flow coming from north of the Wahpool divide (Q_n): $nf = 11$, and $T = 70 \text{ m}^2/\text{d}$.

Therefore, $Q_n = 770 \text{ m}^3/\text{d}$.

Therefore the regional groundwater flow into Lake Tyrrell, ($Q_s + Q_n$), is

$$(Q_s + Q_n) = 1,100 \text{ m}^3/\text{d} \tag{12.4}$$

However, there are also concentrated brines flowing from Lake Timboram, which pass beneath the Wahpool groundwater divide. An estimate based on salt balances suggests that about 33% of the inflow into the Wahpool—Timboram system leaks into Lake Tyrrell (see Chapter 12.3.1). This gives an additional groundwater inflow of $800 \text{ m}^3/\text{d}$, resulting in a total inflow into Lake Tyrrell of $1,900 \text{ m}^3/\text{d}$.

The inflow into Tyrrell of $1,900 \text{ m}^3/\text{d}$ is surprisingly less than the $2,400 \text{ m}^3/\text{d}$ passing into the Wahpool-Timboram sink. The latter, although considerably smaller in area than Tyrrell, lies directly across the flow path, and therefore causes a significant diversion of groundwater that would have otherwise fed Lake Tyrrell. It is notable that the value of $1,900 \text{ m}^3/\text{d}$ is about 40% of the value obtained if the calculation is carried out on the data close to the lake edge without any corrections for interface effects, density variations, etc.

12.3.3 Salt Input into Lake Tyrrell

12.3.3.1 Input from groundwater sources

An assessment of the salt input into Lake Tyrrell can be obtained from the groundwater inflow data by splitting the total inflow into its components. These are the $1,100 \text{ m}^3/\text{d}$ of regional groundwater having a salinity of 40,000 mg/l and $810 \text{ m}^3/\text{d}$ of Timboram-Wahpool reflux brines having a salinity of 120,000 mg/l.

The total daily salt input to Lake Tyrrell from these two sources is therefore:

$$\begin{aligned}
 &141.2 \text{ tonne of salt/day} \\
 &= 5.2 \times 10^4 \text{ tonne of salt/year} \\
 &= 2.4 \times 10^4 \text{ m}^3/\text{year (based on a S.G. for halite of 2.15)}
 \end{aligned}
 \tag{12.5}$$

Apart from the radial flow into the lake from surrounding regions, there is also an unspecified inflow of Tyrrell reflux brines returning to the lake. This inflow is attributed to small residual pressures within the reflux brines, which are of recent origin (Chapter 12.14).

While they are not considered a significant contributor to the overall salt budget, any influence they may have will be to lessen the time taken for the development of the salt crust now found on the lake floor.

12.3.3.2 Input from non-groundwater sources

Additional salt inputs into the Tyrrell Basin come from atmospheric salts in rainwater and salts brought in during the occasional flood flow in the Tyrrell and Lalbert creeks (Table 12.2). These inputs are documented in Macumber (1983a), where it is shown that their present-day

contribution is insignificant alongside that of the groundwater input and was only a minor part of the salt budget in the past.

TABLE 12.2 - Tyrrell Creek and Lalbert Creek waters, 1981

	Tyrrell Ck. 5/2/81 mg/l	Tyrrell Ck. 18/8/81 mg/l	Tyrrell Ck. 19/8/81 mg/l	Lalbert Ck. 19/8/81 mg/l
TDS	40,980	2,459	985	233
Chloride	20,800	1,300	490	59
Sulphate	5,120	206	63	14
Bicarbonate	Nil	56	55	76
Calcium	320	34	22	15
Magnesium	1,440	76	24	7
Sodium	13,100	754	284	40
Potassium	100	16	31	6
Silica	20	11	14	12
Bromide	73			
pH	3.4	7.2	7.5	7.6

Table 12.2 gives analyses from the Tyrrell Creek—under normal base flow conditions, analysis 1, and as the flood waters reach the lake, analyses 2 and 3. The fourth analysis is from the adjacent Lalbert Creek which was also flowing in August 1981. (Lalbert Creek flows into Lake Timboram.)

12.3.4 Accumulation Time for the Lake Tyrrell Salt Crust

Since the present rate of salt input into Lake Tyrrell is known, an indication of the time it has taken for the accumulation of the 70 mm salt crust on the floor of the lake can be obtained. It is assumed that the lake is closed to outseepage as is shown in Chapters 12.12.4 and 12.14. The calculation below also provides some indication of the time during which the present closed lake regime has been in existence.

Calculation:

$t = V/v$, where t is the time taken to form 70 mm salt crust (years)

V = the volume of salt on the lake floor = area of lake floor $\times$ salt thickness

$V = 1.2 \times 10^8 \times 7 \times 10^{-2} \text{ m}^3$

v = the annual rate of salt accumulation = $2.4 \times 10^4 \text{ m}^3/\text{year}$

$t = 350 \text{ years}$

(12.6)

The salt budget therefore suggests that the present hydrological status of Lake Tyrrell has been in existence for only about 350 years, with a salt accumulation rate of 50 years/cm.

Therefore, on the basis of the present flow system, which assumes no significant outseepage, it has taken only about 350 years to accumulate the salt crust in Lake Tyrrell. Even in the unlikely event of the chosen transmissivity values being a factor of 2 in error, this still gives a relatively short time span of from 180 to 700 years for the salt crust to form. The calculation, although requiring qualifications with respect to transmissivity values, nevertheless gives clear proof that the present hydrological status of the lake (that of a salt accumulator) is only a recent innovation. It also shows that the bulk of the salt that entered the lake in the past has been subsequently lost from the lake floor.

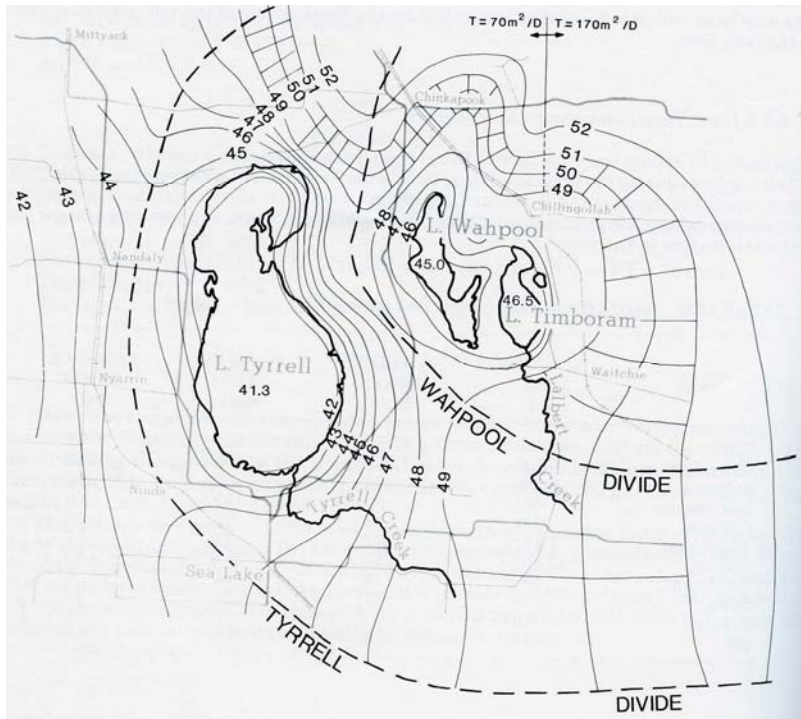


Figure 12.7 - Flow net analysis for the Tyrrell Basin

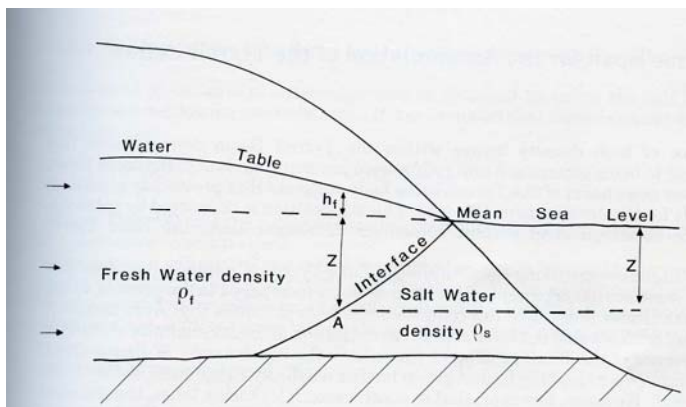


Figure 12.8 - Notation for Ghyben-Herzberg equilibrium

If the same rate of salt inflow had occurred for the past 16,000 years (that is, since the cessation of the last significant phase of lake floor deflation and consequent salt loss), and if there had been no outseepage through the floor, then a salt layer about 3.3 m thick would have accumulated on the lake floor. While it is absent from the lake surface, it is shown (Chapter 12.3.6) that this volume of salt comprises much of the salt stored in the aquifer beneath Lake Tyrrell as reflux brines.

12.3.5 Temporal Variations in the Salt Inflow into Lake Tyrrell

A limit to the rate of salt inflow into the Tyrrell Basin and hence into Lake Tyrrell from groundwater sources can be demonstrated, even though major climatic fluctuations may occur. (A more detailed account is given in Macumber, 1983a.) This limit arises from the restricted capacity of the

Parilla Sand aquifer to transport groundwater; it is imposed by the low average gradients of the land surface and by the limits of aquifer transmissivity. Taken together, these variables provide an upper limit for the rate of groundwater flow towards the Tyrrell Basin, which could at most be only 20% greater than under the present-day regime and operate over limited time periods. Therefore, the calculated value of about 3.3 m thickness of salt per square metre area of lake floor, shown as entering the lake from the groundwater system over the last 16,000 years, although based on present-day inflow rates, would not be significantly changed even with additional input during periods of higher water budgets known to have occurred as a consequence of climatic variations during the Holocene.

12.3.6 The Time Span for the Accumulation of the Tyrrell Reflux Brines

While the presence of high density brines within the Tyrrell Basin demonstrates that conditions conducive to brine generation and reflux were present from late Pleistocene times, the absence of brines downbasin of the Tyrrell brine body suggests that previously conditions were not favourable for the development of brines. This supposition is supported by palaeohydrological evidence based on stratigraphic sequences developed along the Lake Tyrrell shoreline.

Well developed high beaches along the shoreline of Lake Tyrrell (Chapter 3.3.1) show that prior to ca. 32,000 years BP. the lake held about 14 m of water (compared to the present winter level of about 0.5 m). Under these lake-full conditions the high salinities that were necessary to produce the reflux brines would not have arisen. An indication of the lake salinity at the time is given by the presence of the small gastropod, *Coxiella* in the beach sands. Williams (1981) places *Coxiella striata* in a halophilic faunal group having a salinity range from about 10,000 mg/l up to 60,000 mg/l. He notes, however, that in southwestern Victorian lakes, the observed salinity range for *Coxiella striata* is from about 3,000 mg/l up to 50,000 mg/l. While it is not known how long the high lake levels were in existence at Lake Tyrrell, it is thought that the high water phase may have lasted for at least 20,000 years.

Whatever the period of time it was in operation, it is clear that recharge from the lake into the groundwater systems during this high lake phase would have been significantly fresher than that entering as reflux brines over the last 32,000 years. During the lake-full period, groundwater outflow from the Tyrrell Basin would have been facilitated by higher water tables downslope of Lake Tyrrell, since an increase in saturated aquifer thickness by about 14 m (based on a 14 m higher lake level) would incorporate into the flow system the more permeable upper unit of the Parilla Sand, the base of which is at present slightly above the water table level in the zone to the west of the lake (Figs 2.29 and 2.30).

Under the high lake level regime, any brine body previously formed and still present within the Tyrrell Basin would be gradually flushed downbasin and replaced by fresher recharging lake waters. It is therefore likely that certain of the groundwaters lying to the west of Lake Tyrrell beyond the limits of the Tyrrell reflux brines have their origins as recharge lake waters dating to the period of higher lake levels.

Once the lake levels fell, a lunette began to form at Lake Tyrrell, showing that the lake had dried out - with conditions being even drier, at least seasonally, than today. Lake levels continued to fluctuate from this time; however, it is unlikely that the lake was more than 3 m to 4 m deep at even the wettest times during the late Pleistocene or Holocene. During this period conditions conducive to the formation of very saline lake brines were present intermittently, and under favourable hydrostatic conditions these brines migrated into the groundwater system to become the Tyrrell reflux brines.

Additional information on the time taken to accumulate the reflux brines can be obtained by using the present-day rate of accumulation of salt in the Tyrrell Basin. While there was clearly loss from the lake surface by deflation during the two lunette forming periods terminating at about 16,000 BP, even during these times the bulk of the salt was stored in the aquifer and was largely

protected from deflationary processes. Two approaches have been adopted, each being a check on the other: firstly, to determine from the Lake Tyrrell sub-basin alone the time taken to accumulate the Tyrrell reflux brines; and, secondly, to take the entire Tyrrell Basin and find the time taken to accumulate the salt stored in both the Tyrrell and Timboram–Wahpool reflux brines.

12.3.6.1 Estimate of the time taken for the accumulation of the Tyrrell reflux brines

An indication of the rate of outseepage can be obtained by using the salt balance between inflowing and outflowing groundwaters. If it is assumed that there is no net salt accumulation in the lake, then:

$$Q_i C_i = Q_o C_o$$

where Q_i and Q_o are the rates of inseeping and outseeping groundwater, and C_i and C_o are their respective concentrations.

The average salinity of the inflowing groundwaters is about 74,000 mg/l, consisting of regional groundwaters (40,000 mg/l) and Timboram-Wahpool reflux brines (120,000 mg/l). The total groundwater inflow into Lake Tyrrell is $1.9 \times 10^3 \text{ m}^3/\text{d}$. The outflowing waters have a salinity of about 250,000 mg/l (the salinity of the Tyrrell reflux brines).

The rate of outseepage, Q_o , is:

$$\begin{aligned} Q_o &= (1.9 \times 10^3 \times 7.4 \times 10^4) / (2.5 \times 10^5) \text{ m}^3/\text{d} \\ Q &= 5.6 \times 10^2 \text{ m}^3/\text{d} \end{aligned} \quad (12.7)$$

Therefore, a reflux rate of $5.6 \times 10^2 \text{ m}^3/\text{day}$ is suggested. If the volume of Tyrrell reflux brine beneath the lake is known, then the time (t) taken for its accumulation can be found by:
t = brine volume in storage in aquifer / rate of reflux into aquifer

$$t = (A \times b \times p) / Q \quad (12.8)$$

t is 24,500 years (maximum) or 6,100 years (minimum)

where A is area of basin underlain by Tyrrell reflux brines ($1.8 \times 10^8 \text{ m}^2$)
b is aquifer thickness (70 m)
p is porosity (0.4, based on the data of Morris and Johnson, 1967, as uppermost estimate, or 0.1 as lowermost estimate),

This calculation suggests that the Tyrrell reflux brines have accumulated relatively recently, certainly within the last 32,000 years, when conditions conducive to brine formation have existed.

The result must be only approximate since it assumes salt inflow only from the groundwater source and steady state inflow rates through time. Departures from these assumptions will work towards reducing the actual time over which the brines accumulated. To accommodate this, the higher salinities in the range of average salinities for inflowing groundwaters are used in this calculation. Were the lower salinities used (100,000 mg/l for the Timboram-Wahpool brines and 35,000 mg/l for the regional groundwaters), then periods of 28,000 years and 7,000 years respectively would have been obtained. However, the uncertainties do not change the validity of the conclusion that the volume of reflux brine stored beneath Lake Tyrrell can be readily accounted for by reflux processes operating within the last 32,000 years.

An additional check on these calculations can be made by comparing the present rate of salt inflow into the entire Tyrrell Basin (that area including lakes Timboram and Wahpool) with the

volume of salt stored in both the Tyrrell and Timboram-Wahpool reflux brines. The total volume of salt stored in the Timboram-Wahpool and Tyrrell reflux brines of the Tyrrell Basin was calculated as being between $6.9 \times 10^8 \text{ m}^3$ (maximum using a porosity of 0.4) and $1.7 \times 10^8 \text{ m}^3$ (porosity of 0.1)—for details see Macumber (1983a).

The annual salt inflow into the entire basin is $2.4 \times 10^4 \text{ m}^3/\text{year}$, making the time taken for the brine body to form as being between 28,000 years and 7,000 years. This compares favourably with the respective values of 24,000 years and 6,100 years obtained for the Tyrrell sub-basin alone.

12.3.7 Brine Accumulation beneath Lake Tyrrell-Summary

Since the water table divide to the west of Lake Tyrrell suggests that the lake is closed to outseepage, the conditions within the lake basin should be conducive to the accumulation of a significant thickness of salt by the process of evaporative concentration of the lake waters. While the larger part of the lake water is derived from direct precipitation and surface runoff, only a small part of the lake water comes from inflowing groundwater. By contrast, however, the bulk of the salt entering the lake is provided by groundwater inflow.

Flow net analysis of the groundwater regime in the vicinity of Lake Tyrrell indicates an average contemporary inflow of $1.9 \times 10^3 \text{ m}^3/\text{day}$, which gives an annual salt cover of 0.2 mm over the lake floor. The analysis and dating of the late Quaternary shorelines shows that the lake has not been subject to surface overflow or flushing within the last 32,000 years. Therefore, at an accumulation rate of 200 mm/1000 years, a salt layer 2 m thick should have been deposited on the lake floor from groundwater inflow in the Holocene period alone.

However, the average thickness of salt on the lake floor is only 70 mm, indicating that the lake has lost almost all the salt that has entered - the 70 mm on the lake floor represents the accumulation of only about 350 years of groundwater inflow at present-day inflow rates.

While salt is missing from the lake floor, the entire aquifer beneath the lake is saturated with water having an average salinity of about 250,000 mg/l. The volume of salt stored beneath the lake accounts for the bulk of the salt inflow into the lake basin over the last 32,000 years, during which period, on the basis of palaeohydrological data from the Lake Tyrrell shoreline, conditions existed intermittently for the formation of reflux brines. The palaeohydrological analysis, together with the flow net analysis and salt budget calculations, underpin the more general conclusions that a reflux process has formed the groundwater brines within the basin, and that this process has occurred within the last 32,000 years. The total area covered by the reflux brine body is about 350 square kilometres.

12.3.8 Brine Accumulation beneath Other Mallee Lakes

The situation at Lake Tyrrell is not unique - similar brine occurrences have been found beneath many of the larger lakes and boinkas of northwestern Victoria (Macumber, 1980). Chemical analyses of these waters show them to be essentially concentrated versions of the regional groundwater in the Parilla Sand aquifer, which have undergone evaporitic concentration in the discharge lakes prior to their return (or reflux) to the groundwater regime (Table 8.1). A throughflow process is indicated. The high salinity of the refluxed waters stems from the high evaporation rates that exist in the area. Where playa lakes in northern Victoria undergo an annual drying cycle, the lake waters are rarely less saline than the regional groundwaters and are commonly two to ten times more saline than the groundwaters.

While the loss of salt from the lake floor into the underlying aquifer is readily understood where throughflow conditions exist in the absence of a groundwater divide (such as at Lake Timboram), it can be shown that in certain instances throughflow may readily occur even where a significant divide is present. This situation may arise in areas such as Tyrrell and the Raak and Pink Lakes

boinkas, where high salinity former lake brines are found replacing less saline regional groundwater in the underlying aquifers.

12.4 The Influence of Variable Density Waters on Groundwater Flow

12.4.1 The Surface Capture of Regional Groundwater Flow in the Vicinity of Large Discharge Lakes - Ghyben-Herzberg Interfaces in Continental Settings

The presence of high density brines beneath the lakes strongly influences the regional groundwater flow system near the lakes. The large density differences between the groundwaters (the density of the Tyrrell brines is 1.16 g/cm^3 , while that of the Timboram-Wahpool brines is 1.09 g/cm^3 , and the regional waters 1.03 g/cm^3) result in the development of a Ghyben-Herzberg interface (Ghyben, 1888; Herzberg, 1901), which, in turn, forces the lower density regional groundwaters towards the surface, ensuring their total capture by the lake systems. The process is analogous to that occurring in coastal aquifers and is the reverse of that occurring beneath ocean islands. With the high density brines infilling the aquifer, there can be no underflow of regional waters beneath the lake, as is the case with the throughflow lakes and the uniform density groundwaters described by Winter (1976, 1978).

Once established, the system is self-perpetuating, since the low groundwater gradients do not permit sufficiently high hydraulic heads immediately upbasin of the lake, which are necessary for any flushing of the aquifer by the process of downbasin groundwater flow alone. The destruction of the system can only be achieved by the imposition of continuously high fresh water lake levels of sufficient depth to overcome the closing influence of the groundwater divide. The fresh water may then replace the existing denser groundwater brine, which is in turn flushed downbasin and out of the system. The latter condition requires that evaporitic concentration of lake waters be kept to a minimum with no seasonal drying out. These conditions (high lake levels and low salinities) last existed in the Tyrrell Basin prior to 32,000 years BP. when Lake Tyrrell held 14 m of water.

12.4.2 High Density Reflux Brines, the Groundwater Divide, and the Outflow Potential of Lakes in Terms of the Ghyben-Herzberg Concept

Once the aquifer underlying a lake is filled with high density brines, the throughflow character developed at the time of initial outseepage is enhanced. This occurs despite the presence of the groundwater divide, since the significantly denser brines may generate sufficient hydraulic pressure deeper within the aquifer to overcome the hydrostatic pressures produced at the divide. Conversely, closure of the lake to outseepage is achieved when the groundwater divide is sufficiently high above the lake to ensure that hydrostatic pressures produced at the base of the aquifer beneath the divide are greater than the countering pressures produced by the denser brines within the aquifer beneath the lake. Where the reflux brines are essentially static, the hydrostatics may be considered in terms of the Ghyben-Herzberg approximation developed for coastal aquifer systems.

The existence of a salt water wedge and its accompanying interface in a coastal situation arises from the density differential between the inflowing groundwaters and the salt water occupying the aquifer on the seaward side of the shoreline. The Ghyben-Herzberg equation for the depth to the interface between the salt and fresh waters in a lacustrine setting is:

$$Z = (\rho_f H_f) / (\rho_s - \rho_f) \quad (12.9)$$

$$\text{or } Z = (\rho_f H_f / \rho), \text{ where } \rho = (\rho - \rho_f) \quad (12.10)$$

where Z is the depth to the interface below the lake floor datum
 H_f is the height of the water table above the lake datum
 ρ_f is the freshwater density - the regional groundwater
 (S.G. 1.03 for this study)
 ρ_s is the saltwater density - the reflux brines
 (S.G. 1.16 for the Tyrrell reflux brines)

The use of the Ghyben-Herzberg concept to equate pressures across the interface results in an approximation for the depth of the interface below the surface, the equation being in terms of water densities and fresh water head. While Ghyben-Herzberg interfaces are generally considered in terms of coastal settings, they also have application in a continental setting, as occurs in the salt lake province of northwestern Victoria.

For example, at the Tyrrell groundwater divide H_f is about 3.7 m (45 m AHD) and the calculated depth to the interface using average values for densities is 29.3 m (15.7 m AHD) - by equation 12.10.

$$Z = (1.03 \times 3.7) / (1.16 - 1.03) = 29.3 \text{ m}$$

This result indicates a fall of about 8 m in the elevation of the interface for each metre rise in the water table above the lake datum. For comparison, the sea water - fresh water equivalent is a 40 m fall in elevation for each metre rise in water table above sea level. The considerable difference in depth to the interface per metre rise in water table height reflects the very much greater density differences that exist in the salt lake environment compared to those in a coastal situation.

Important hydrologic differences exist between the continental setting and a coastal setting, since in the latter instance, the salt water intrusion eventually wedges out on passing inland in response to the gradually rising water table. In the case of an inland lake bounded on its downbasin side by a groundwater divide, the salt water intrusion may also wedge out at some point between the lake shoreline and the divide. However, should the salt water intrusion reach as far as the groundwater divide, where the fresh water head downbasin of the lake is at its highest value, then any further intrusion cannot be contained, since beyond this point the fresh water head declines as the water table slopes downbasin. Should this occur, the salt water underflows the fresh water at the divide and escapes from the basin, probably mixing with regional waters by upconing. Since the requirement for leakage is that the theoretical position of the interface (point A, Figure 12.9) is above the base of the aquifer (point B) at the divide, then, for a 3.7 m divide as occurs at Lake Tyrrell, the aquifer must extend to at least 29.3 m below lake level for the lake basin to be open. This example brings forth a general conclusion that thin aquifers and high groundwater divides are conducive to closed lakes, while thick aquifers favour outseepage. Outflow may also be enhanced under two quite different hydrological conditions:

Firstly, it will be favoured under more arid conditions, when the regional water table and the groundwater divide are low. (The actual outflow is here at a minimum, since surface water budgets are low.)

Secondly, where a closing divide exists, a rise in lake level with a consequent increase in hydrostatic head in the vertical brine column beneath the lake may also lead to outseepage. An indication of the influences of aquifer thickness, groundwater divide height, and reflux brine density on lake outseepage is given in Figure 12.10. This shows the depth below mean lake level to the interface between reflux brines (of varying densities) and the regional Parilla

Sand groundwaters (of average density 1.03 g/cm^3) for a given height of the groundwater divide above mean lake level. The depths of the base of the Parilla Sand aquifer beneath the groundwater divide at Lake Tyrrell and immediately to the west of Lake Timboram are 34 m and 50 m below the floors of the respective lakes.

At Lake Timboram, where the reflux brines have a density of 1.09 g/cm^3 , a groundwater divide rising more than 2.8 m above the lake floor is required to prevent outseepage. At Lake Tyrrell, where the reflux brines have a density of 1.16 g/cm^3 , a divide of about 4.2 m is needed to prevent outflow. There is, however, no significant groundwater divide at Lake Timboram to impede outseepage, while at Lake Tyrrell the divide height of 3.7 m above the lake floor is 0.5 m below that theoretically required to prevent outseepage.

However, the potential for outseepage at Lake Tyrrell is not as great as that suggested by the above calculation, since the actual groundwater density at the divide is slightly greater than the average regional groundwater density. As is shown below, the additional increase in regional groundwater density at the divide is such that if the groundwater system were to conform to the present-day hydrostatic conditions determined by the elevations of water tables and lake levels and by groundwater densities, then the lake-groundwater system at Lake Tyrrell would lie virtually at the threshold of outseepage during summer months, and hydraulic gradients conducive to outseepage would arise in winter. The sensitivity of the lake hydrology to small fluctuations in the height of the groundwater divide is readily apparent from Figure 12.10; for instance, in the case where reflux brines of density 1.09 infill the aquifer, and the groundwater divide falls from 3 m to 2 m above the mean lake level, the interface rises from a depth of 51 m below the lake floor to 35 m below the lake floor. Given an aquifer thickness of 50 m (as is the case for Lake Timboram), then the lake would pass from a state of being closed to one where the interface would be some 15 m above the aquifer base at the divide, and therefore well open to outseepage. A similar effect is produced by raising the mean lake level 1 m. Given the above responses, the overall hydrology of the closed lakes in the semi-arid regions of northern Victoria is one of metastable equilibrium, in which small changes in water budget may lead to an enhanced throughflow capacity of the lakes, thereby promoting salt loss from the lake floor to the underlying aquifer prior to it being flushed out of the lake basin. These processes are put forward as an explanation for the absence of halite evaporite deposits in most northern Victorian salt lakes.

12.4.3 The Influence of the Tyrrell Fault on the Outflow Potential of the Tyrrell Basin

When regional groundwater densities are used in equation 12.10, the height of the groundwater divide is seen to be insufficient to prevent outseepage under the present day hydrologic regime. However, salinity measurements from the Bimbourie 1 bore, situated close to the divide (Figure 2.34), show the waters to be slightly more saline than the average regional groundwaters, with a density of about 1.045 g/cm^3 . This bore is about 5 km west (downbasin) of the Bimbourie 2/3 piezometer nest (Figure 12.17), where high salinity reflux and transitional waters (100,000 to 250,000 mg/l) were found over much of the central and lower sequences (Figure 12.13). Instead, in the Bimbourie 1 bore, salinities ranged from about 52,000 mg/l at the water table to about 57,000 mg/l towards the base of the Parilla Sand aquifer (Table 12.3).

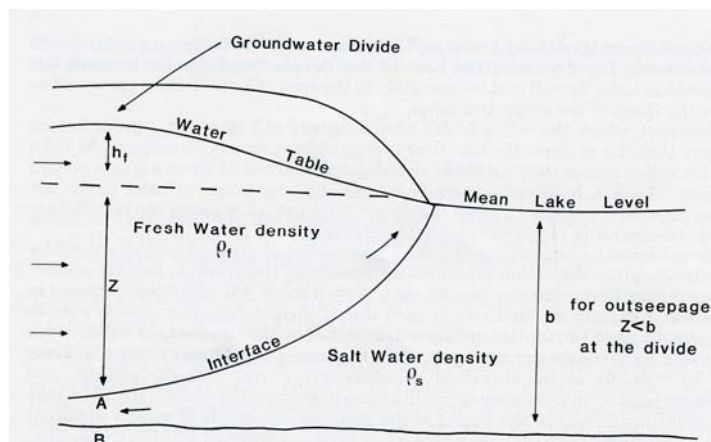


Figure 12.9 - Salt water outseepage beneath a groundwater divide

TABLE 12.3 - Salinity and density variations in the Bimbourie 1 bore

Depth (m)	Salinity (mg/l)	Density (g/cm ³)
25- 30	53,000	1.042
30- 40	54,500	1.045
40- 50	56,500	1.046

Depth to water table is 20 m

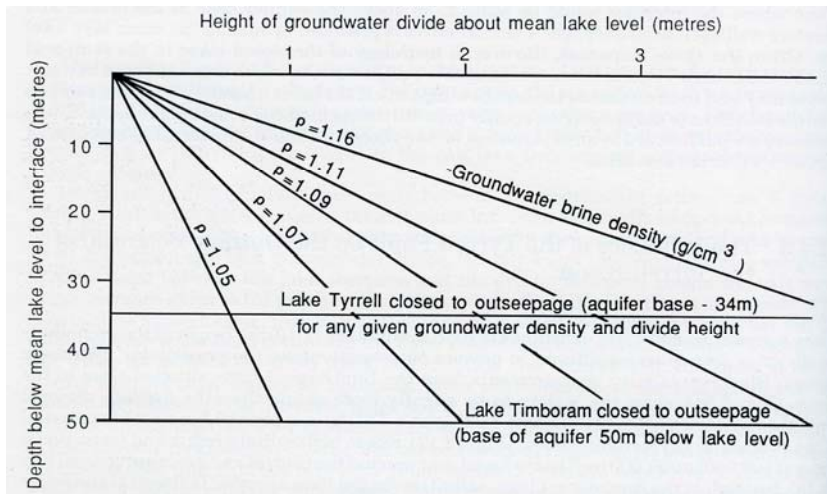


Figure 12.10 - Graphical expression of the varying influences of the height of the groundwater divide and the groundwater density on the outseepage potential of lakes underlain by reflux brines under a Ghyben-Herzberg equilibrium.

The fresh water density is taken as 1.03 g/cm³. The influence of aquifer thickness on outseepage potential for any given divide height and water density can be seen from the examples of Lake Timboram and Lake Tyrrell, although in reality other factors may also play a decisive role.

The Ghyben-Herzberg interface marks the point at any given depth within the aquifer of pressure equalization between the reflux brines and the regional groundwater. For the closure of the lake basin by equation 12.10 - which implies no movement in the reflux brine - the depth to the base of the aquifer beneath the divide (Z_b) must be less than that required for the brines to underflow the divide (Z). That is, by equation 12.9:

$$Z = (\rho_r H_f) / (\rho_s - \rho_r) > Z_b$$

If the floor is taken as the zero datum, the density differential as 0.115 ($\rho_s = 1.16$ and $\rho_r = 1.045$), and the height of the groundwater divide as 3.7 m (45.0 m AHD) above the lake floor (41.3 m AHD), then closure occurs when:

$$\begin{aligned} (1.045 \times 3.7) / 0.115 &> Z_b \\ \text{or } Z_b &< 33.6 \text{ m} \end{aligned} \quad (12.11)$$

Therefore, for closure to outleakage, the base of the aquifer at the divide must be within 33.6 m of the lake floor. This is at an elevation of 7.7 m AHD. Beneath Lake Tyrrell the base of the Parilla Sand is at an elevation of about -20 m AHD, however, the Parilla Sand is downwarped to the west of Lake Tyrrell, and at the groundwater divide the base of the aquifer is considerably higher

than beneath the lake. Without the warp, the base of the aquifer at the groundwater divide would have been about 27 m below the interface, providing a deep and wide gateway for outflow. However, data from the Bimbourie 1 bore shows that the base of the aquifer on the high side of the structure is at an elevation of about 7 to 8 m AHD (see gamma log, Figure 12.11). The elevation of the base of the aquifer at the groundwater divide is therefore virtually that at which the basin closes (or opens).

The Tyrrell Basin is thus near to the threshold between outflow and closure, the balance clearly being very delicate and requiring only minor changes in basin hydrology to push the equilibrium either way.

12.4.4 Assumptions in the Application of the Ghyben-Herzberg Equation

The use of the Ghyben-Herzberg equation for interface calculations assumes horizontal flow in the fresh water; that is, there is no head loss between the water table and the interface. It implies that at the divide a vertical equipotential exists from the water table to the interface. The extent to which the groundwater flow system deviates from this situation ultimately determines the actual hydrostatic head opposing outflow. Where there is no deviation, the closing head is that of the water table elevation at the divide. A check on the applicability of the Ghyben-Herzberg equation for the Tyrrell situation can be made by comparing environmental-water heads - a measure of vertical head (see Chapter 12.9) - in the Bimbourie 1/SR 6032 piezometer nest situated close to the divide. From Table 12.8 it can be seen that the environmental-water head (H_{in}) in Bimbourie 1 (at an elevation of 17 m AHD) is 44.55 m, while that of the SR 6032 piezometer (virtually at the water table) is 44.50 m. These values are almost the same, indicating that equipotentials are essentially vertical; that is, flow within the aquifer is virtually horizontal and normal to the 44.5 m vertical equipotential. The head measurements in the two bores support the earlier contention (Chapter 12.2 and Figure 12.4) that horizontal flow is to be expected close to the Tyrrell groundwater divide where there is a low hydraulic gradient and very little contribution from local recharge. One further limitation to the use of the Ghyben-Herzberg equation is that it requires that no flow occurs in the saline water; that is, the hydrostatic head of the salt water is everywhere at the same elevation as mean sea level (or mean lake level in this study). For this setting, a steady state no-flow regime must be assumed for the reflux brines.

12.4.5 Lake Outseepage but No Basin Outflow of Brine

The calculations based on the Ghyben-Herzberg equation suggest that at present the basin may be just open or just closed to outseepage by underflow of the groundwater divide. Furthermore, the absence of the bulk of the inflowing salt from the lake floor indicates that outflow conditions existed at the lake for considerable periods in the past. Yet the earlier salt balances indicate that the bulk of the salt entering the basin over the last 32,000 years (indeed, commencing at some time between 28,000 and 6,000 years ago) is still retained within the basin, stored within the reflux brines. This conclusion is supported by chemical data from the Bimbourie 1 bore situated near the divide, where no reflux brine was detected in the section. An explanation for the retention of the reflux brines within the Tyrrell Basin, compatible with the earlier contention that the lake basin may be readily opened to outseepage as a consequence of small increases in lake level or by falls in the height of the groundwater divide, is that, while gradients conducive to underflow may exist for prolonged periods, the average rate of downbasin flow has been insufficient for the brine front to advance the 6 km to the groundwater divide over the period during which outflow conditions have prevailed (see Macumber, 1983a, for a more detailed discussion).

12.5 Hubbert's Dynamic Concept

The Ghyben-Herzberg concept gives an indication of the manner in which outseepage may be enhanced by the density differential between the regional groundwaters and the reflux brines. However the concept is based on hydrostatics and assumes no brine movement. In coastal fresh water-seawater systems there is normally not only a significant fresh water movement towards the coast, but also considerable motion in the underlying salt water. In order to deal with the dynamic equilibrium of the mobile fresh water – salt water interface, Hubbert (1940) showed that the depth to the interface (Figure 12.12) could be more accurately defined by the equation:

$$Z_i = (\rho_s H_s) / (\rho_s - \rho_f) - (\rho_f H_f) / (\rho_s - \rho_f) \quad (12.12)$$

where Z_i is the elevation of the interface
 H_f is the static level of fresh water in a bore terminating in fresh water H_s is the static level of saline water in a bore terminating in salt water
 ρ_f is the fresh water density
 ρ_s is the saline water density

The Hubbert equation is commonly expressed as:

$$\rho_f H_f = \rho_s H_s - Z (\rho_s - \rho_f) \quad (12.13)$$

or more generally (where it is applied to two immiscible liquids of density ρ_1 and ρ_2 , where $\rho_2 > \rho_1$, each having hydrostatic heads H_{1p} and H_{2p} respectively), it becomes:

$$\rho_1 H_{1p} = \rho_2 H_{2p} - Z_i (\rho_2 - \rho_1) \quad (12.14)$$

Where the static level of the saline water in equation 12.12 is at sea level, i.e. $H_s = 0$, indicating no flow in the intruded water, then the equation reduces to the Ghyben-Herzberg equation. Hubbert's equation implies a sharp interface; however, there is commonly a varying wide transition zone of intermediate density water between the intruded salt water and the overlying fresh water. A further limitation is that it implies that, as in the Ghyben-Herzberg equation, there is no vertical groundwater flow component.

12.6 The Transition Zone

The concept of a sharp interface is a most useful approximation of the flow regime, especially for mathematical manipulation; in practice, relatively sharp changes from regional water to reflux brines have been only occasionally noted in the Tyrrell Basin. In most cases the interface is replaced by a transitional zone ranging in thickness from a few metres to over twenty metres. This is readily seen in piezometers situated on the Tyrrell shoreline (Figure 12.13).

In instances where the interface is stable, a transitional zone develops in response to mechanical (or convective) dispersion and molecular diffusion. The first process arises from velocity variations of the flow within the intergranular pore spaces; at the same time, lateral variations in salinity produce concentration gradients, which in turn result in particle flux from regions of high to low salinity. In general, the magnitude of dispersion for uniform sands is only a few metres over a travel distance of 10^3 m (Todd, 1980). The one-dimensional form of the dispersion differential equation, where the invading fluid enters a porous medium on a broad front, is (Bruch and Street, 1967)

$$D_1 \frac{\partial^2 C}{\partial x^2} - v_x \frac{\partial C}{\partial x} = \frac{\partial C}{\partial t} \quad (12.15)$$

This equation describes the situation for a nonreactive solute in a saturated homogeneous isotropic medium, under steady state flow conditions

where

C	is the solute concentration of the invading fluid
D_1	is the longitudinal dispersion coefficient
v_x	is the average linear flow velocity
x	is the distance along the flow path
t	is the time since the invasion commenced signifies a partial differential.

The longitudinal dispersion coefficient (D_1) can be expressed as:

$$D_1 = \alpha_1 v + D^* \quad (12.16)$$

where α_1 is a characteristic property of the porous medium known as dispersivity, and D^* is the coefficient of molecular diffusion for the solute in the porous medium.

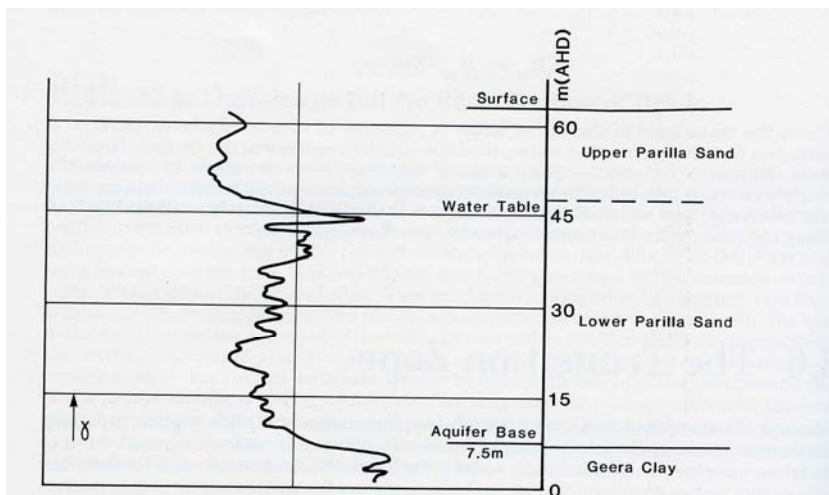


Figure 12.11 - Bimbourie 1 gamma log

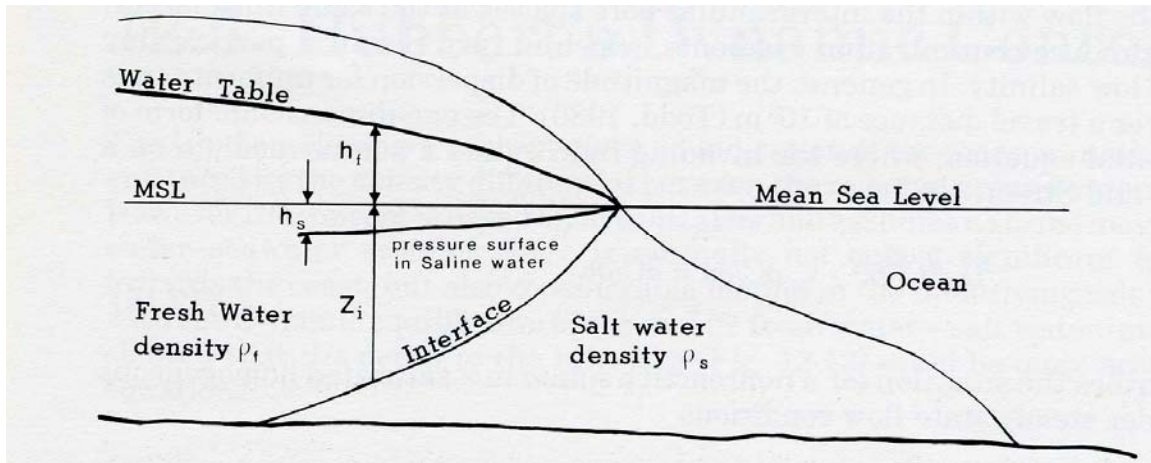


Figure 12.12 - Notation for Hubbert's dynamic concept

Values of D^* commonly range from 5×10^{-9} to $10^{-11} \text{ m}^2/\text{s}$.

Given the following boundary conditions:

$$\begin{aligned} C(x,0) &= 0, & x &\geq 0 \\ C(0,t) &= C_0, & t &\geq 0 \\ C(\infty,t) &= 0, & t &\geq 0 \end{aligned}$$

an analytical solution to the equation is given by Ogata, 1970:

$$C = C_0 0.5(\text{erfc} [(x-v_x t) / 2(D_1 t)^{1/2}] + \exp (v_x / D_1) \text{erfc} [(x + v_x t)/2(D_1 t)^{1/2}]) \quad (12.17)$$

Where C is the solute concentration at distance x along the flow path
 C_0 is the initial concentration of solute in the invading fluid erfc is the complimentary error function = $(1 - \text{erf})$
 D_1 , v , x , and t are as previously defined

At low flow velocities diffusion is the important contributor and $D_1 = D^*$, while equation 12.15 reverts to:

$$\delta C / \delta t = D^* \delta^2 C / \delta x^2 \quad (12.18)$$

This differential equation expresses Fick's second law, which relates the concentration of a diffusing solute to time and space. Values of C at different times can be calculated from equation 12.19, (Freeze and Cherry, 1979).

$$C = C_0 \text{erfc} [x/2(D^* t)^{1/2}] \quad (12.19)$$

where C is the concentration at distance x along the flow path after time t ; all other terms previously defined.

An indication of the influence of diffusion in the transition zone between the Tyrrell reflux brines and the regional waters is given in Table 12.4. In the calculations, values of $10^{-10} \text{ m}^2/\text{s}$ were chosen for D^* , while the concentrations of the brines and regional groundwaters are taken as 250,000 mg/l and 40,000 mg/l respectively.

TABLE 12.4 - Diffusion across the saltwater interface (concentration at distance x after time t in mg/l)

Time (years)	1 m	Distance (m) 10 m	100 m
100	84,000	40,000	40,000
1,000	189,000	40,000	40,000
10,000	230,000	84,000	40,000

Table 12.4 shows that diffusion has a very limited influence beyond the immediate vicinity of the interface.

While under steady state conditions the width of the transition zone may be determined using the hydraulic properties of the aquifer and the coefficient of dispersion, external hydrological factors causing the unsteady displacement of the interface may play the major role in the development of the transition zone. These processes are readily seen in coastal aquifers which are subject to intensive pumping. For instance, at Honolulu-Pearl Harbour, Hawaii, a transition zone more than 300 m thick resulted from concentrated pumping of the coastal aquifer (Visher and Mink, 1964).

With lakes such as Tyrrell, displacement of the interface can be brought about by lake level fluctuations—both seasonal wetting and long-term filling or drying of the lake. In the latter instances, the position and width of the transition zone may provide a clue to earlier groundwater – lake water equilibria. Seasonal fluctuations of lake level will have greatest impact close to the shoreline, where reversals in the direction of vertical gradient accompany the occasional wet period when lake levels are abnormally high; away from the lake the effect is minimal. More permanent lake level changes, such as accompanied a recent drying of the lake, can affect the degree of lake closure, and thus cause extensive migrations of the interface and corresponding alterations to the width of the transition zone, both close to the lake and well away from the lake. A 40 m wide transition zone in the Bimbourie 4/5 piezometer nest situated 1 km to the west of Tyrrell may be largely the result of the latter process (Figure 12.13).

One further influence on the extent of the transition zone is the effect of long-term water table fluctuation, brought about by either climatic change or by vegetation changes. The response of the water table to the removal of the native eucalyptus cover throughout the Mallee has already been noted (Macumber, 1978b, 1980). Since the height of the water table determines the degree of closure of the lake to outseepage, any change in the water table results in a corresponding movement in the interface, and hence a modification of the transition zone.

12.7 Chemical Determination of the Salt Water - Fresh Water Interface

Since the interface between the reflux brine and the regional groundwater is contained within a transition zone, it is necessary, for the purpose of simplifying analysis, to choose a point to represent the interface. In this study the approach of Schmorak and Mercado (1969) is adopted and assumes a one-dimensional dispersion problem with flow taking place in a vertical direction. Here the 50% concentration is taken as representing the salinity at the position of the interface. The relative salinity (SR) is calculated as a percentage by:

$$SR = 100 (C - C_f)/(C_s - C_f) \quad (12.20)$$

where C is the salinity at a given depth within the transition zone, and C_f and C_s are the salinities of the regional waters and the reflux brines respectively (see Todd, 1980). The depth to the interface can be obtained by plotting SR against depth on probability paper and taking the depth of the 50% value.

Examples of the use of chemical data in investigations of the interface come from two piezometer nests along the western shoreline—the SR 78 and 80 nests (locality Figure 12.17). Here, the salinity at the interface was calculated at 150,000 mg/l, based on 4% regional groundwaters and 26% reflux brines. At the SR 78 nest, the deeper bore (depth 16 m) showed only slight dilution, with a salinity of 235,000 mg/l, while the shallow bore (depth only 3.6 m) had a salinity of 144,000 mg/l, indicating a close proximity to the calculated interface. However, at the SR 80 nest 10 km to the south, the deeper bore (depth 10 m with a salinity of 150,000 mg/l) terminated at the interface, while the shallow bore (depth 3.5 m) contained almost regional groundwaters at 43,000 mg/l.

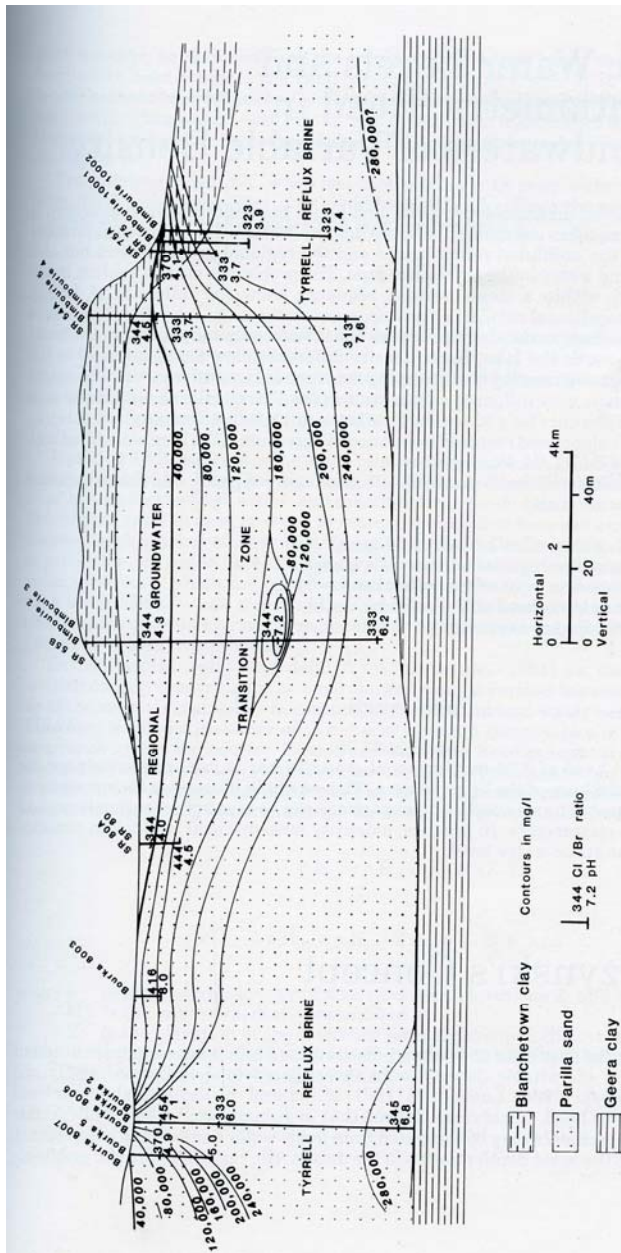


Figure 12.13 - Salinity distribution beneath the western shoreline of Lake Tyrrell (for section line see Figure 12.17)

12.8 Static Water Levels and Potentiometric Head in Groundwaters of Variable Density

The problem of determining potentiometric head, and hence gradients and flow direction, is made more difficult in aquifers containing variable density waters. The density differences exist not only between the undiluted reflux brines and the regional groundwaters but also between these waters and water in the transition zone. The problem is clearly evident in the case where piezometers within a single vertical sequence show the influence of density differences across the transitional zone; it is also important closer to the lake shoreline, where groundwater flow has a strong vertical component as the interface contained within the broad transition zone rises towards the lake. Here, density differences are found not only in the vertical plane but also between nearby bores tapping the same horizontal interval. The extent to which density differences may influence static levels in the Tyrrell lake regime can be seen by comparing the head difference for a 30 m column of brine in a bore bottoming in reflux brines of S.G. 1.17 and the equivalent head required to balance the pressure if the bore were filled with regional groundwater of S.G. 1.04. For instance:

The pressure at the bottom of the 30 m column of saline water is $\rho_s g H_s$, and this is balanced by the fresh water pressure $\rho_f g H_f$.

Where ρ_s is the density of the reflux brines-1.17 g/cm³
 ρ_f is the density of the regional water-1.04 g/cm³
 H_s is the potentiometric head of the reflux brine—30 m H_f is the potentiometric head of the regional water
 g is the acceleration due to gravity

At equilibrium

$$\rho_s g H_s = \rho_f g H_f \quad (12.21)$$

$$\text{i.e. } H_f = 33.75$$

Thus, an additional head of 3.75 m of regional groundwater is required to balance the pressure developed at the base of the brine column. This example illustrates the necessity to apply density corrections before any realistic comparisons can be made of potentiometric data from the Tyrrell Basin piezometers. In practice, a density measurement was taken simultaneously with that of the static water level.

12.9 Luszczynski's Concept

Methods for computing the gradients and velocity distribution for flow between measurement points in groundwaters of variable density were investigated by Luszczynski (1961) and Luszczynski and Swarzenski (1966). Luszczynski (1961) introduced the terms fresh-water head and environmental-water head. Luszczynski showed that in groundwater systems of variable density, the horizontal gradients may be computed from fresh-water heads along a horizontal, i.e. in bores screened to the same depth relative to the datum; the vertical hydraulic gradients must, however, be obtained from the environmental-water heads measured along a vertical. Point-water head refers to the head (or static level) in a bore filled with water of the same density as that found at the screened interval (the screened interval is seen as a point at a given depth within the aquifer)—see Figure 12.14a. Fresh-water head is used for the head of fresh water required to balance the pressure existing at the point, after the bore is filled with fresh water—see Figure 12.14b.

The pressure at point 'i', when the bore is filled with point water, is given by $\rho_i g(H_{ip} - Z_i)$; when filled with fresh water it is given by $\rho_f g(H_{if} - Z_i)$. From the equality of pressures, the relationship between fresh-water and point-water head is:

$$\rho_f H_{if} = \rho_i H_{ip} - Z_i(\rho_i - \rho_f) \quad (12.22)$$

where H_{if} is the fresh-water head at the point 'i' in groundwater of variable density
 H_{ip} is the point-water head at 'i'
 Z_i is the elevation of 'i', with measurements being positive upward
 ρ_i is the density of the point water at 'i'
 ρ_f is the density of the fresh water

Equation 12.22 expresses the same relationship as that in Hubbert's equation for the salt water – fresh water interface (equation 12.13). As in the earlier situations, a density of 1.03 g/cm^3 is used for the fresh water, and therefore fresh-water heads are all in terms of this value. This higher value is used in order to retain compatibility between any calculated values and the very large number of measured values from the Mallee, in general, and the Tyrrell Basin, in particular. Its use is indeed necessary when calculating the theoretical depths of the salt water interfaces, which are between regional groundwaters and reflux brines, not between fresh water (density 1.00 g/cm^3) and reflux brines. This is also the case when considering the effectiveness of basin closure by regional groundwaters, where a fresh water density value of 1.00 g/cm^3 is also inappropriate.

Environmental water was defined by Luszczynski (1961) as 'the water of constant or variable density occurring in the environment along a vertical between a given point and the top of the zone of saturation'. In essence, the environmental-water head is the head produced if the bore is filled with water of variable or constant density, in a manner equivalent to its occurrence within the aquifer. Environmental-water head cannot therefore be readily measured in the field from a simple piezometer; it must be calculated from other head and density data. From the equality of pressures in Figure 12.14, the relationships between point-water and fresh-water heads and environmental-water head is given (Luszczynski, 1961) by:

$$\rho_f H_{if} = \rho H_{ip} - (\rho_f - \rho_a)(Z_i - Z_r) \quad (12.23)$$

$$\rho_f H_{in} = \rho_i H_{ip} - Z_i(\rho_i - \rho_a) - Z_r(\rho - \rho_f) \quad (12.24)$$

where ρ_a is the average density of water between elevations Z_r and 'i'

H_{in} is the environmental – water head at 'i'
 Z_r is the elevation of the reference point from which the average density of water to
 'i' is determined, and above which the water is fresh (see Lusczynski p. 4249) Other
 symbols as previously defined.

The reference point from which density measurements are made must be from within the fresh
 water zone, or, where no fresh water exists, from the top of the zone of saturation. In coastal
 aquifers, where the water above sea level is fresh, sea level maybe used as the reference point,
 and Z_r then is zero. The equations thus become:

$$\rho_f H_{in} = P_i H_{if} - Z_i (\rho_f - \rho_a) \quad (12.25)$$

$$\rho_f H_{in} = \rho_i H_{ip} - Z_i (\rho_i - \rho_a) \quad (12.26)$$

where ρ_a is the average density between mean sea level and 'i'.

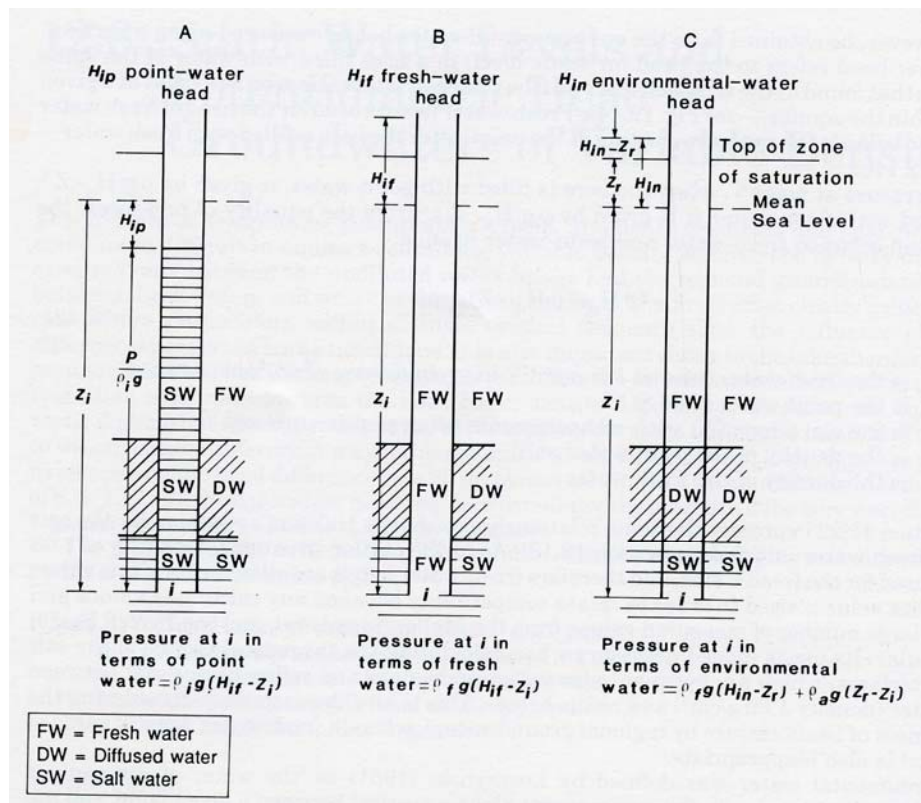


Figure 12.14 - Heads in groundwater of variable density (from Lusczynski, 1961)

In the Tyrrell study, however, while it would have been convenient to make the mean lake level (41.3 m AHD) the density reference point as well as the zero datum for head measurement, the presence of higher density transition zone waters at elevations above mean lake level required that the top of the zone of saturation be used as the reference point for determination of average densities. Equations 12.25 and 12.26 were normally used for all calculations of H_{in} in order to determine the vertical flow component in waters of variable density. The values for average density were, wherever possible, based on data from piezometer nests and other nearby bores. In other instances, estimates were made by averaging the salinity in the bore and that of the top of the zone of saturation.

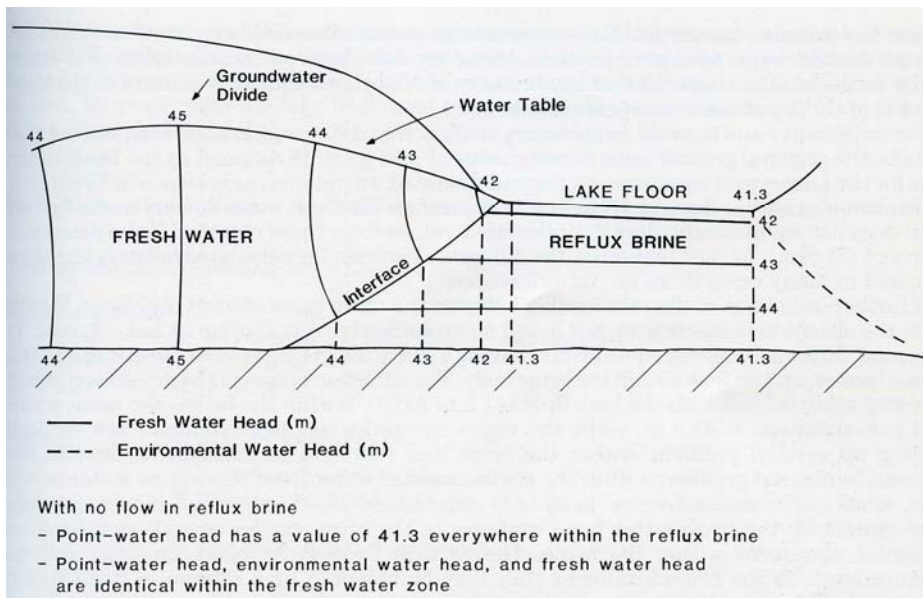


Figure 12.15 - Relationship between potentiometric heads in a groundwater system of variable density in the vicinity of a playa lake. The situation is one where there is no flow in the reflux brine (modelled on Lake Tyrrell).

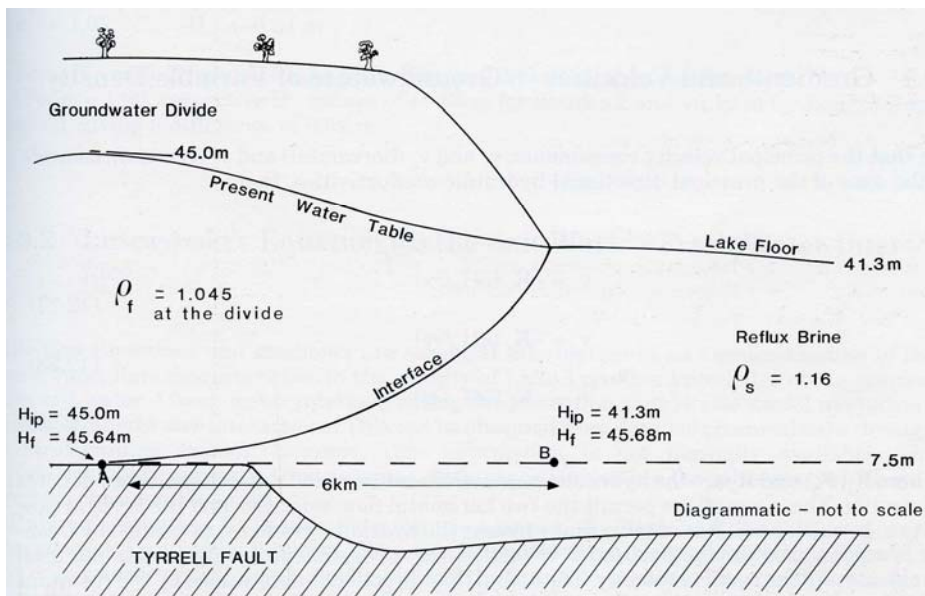


Figure 12.16 - Environmental and freshwater-head measurements in the regional groundwaters at the Tyrrell groundwater divide (point A) and in the Tyrrell reflux brines beneath the lake (point B).

The near equilibrium situation, shown here by the freshwater-heads under lake dry conditions, indicates the delicate balance which exists in the outflow potential of Lake Tyrrell.

For consistency and to avoid unnecessary confusion and the proliferation of equations and symbols, the regional groundwater density value of 1.03 g/cm^3 is retained as the fresh water value for the Lusczynski equations, as previously stated. In practice, as is shown in Table 12.8 and in examples within the text, the use of 1.03 g/cm^3 for the fresh water density in the Tyrrell

Basin does not significantly alter the calculated values from those obtained if the density is taken as 1.00 g/cm^3 . In most instances, the difference between the calculated heads is less than 4 cm, and in many cases there are no differences.

A further indication of the relationship between the three types of head is given in Figure 12.15; the situation is theoretical, but based on groundwater distribution at Lake Tyrrell. It is a typical Ghyben-Herzberg equilibrium, having a sharp interface between the brines and the regional water, and no flow within the brine body. The situation occurs in the dry season, when the water table is level with the lake floor (41.3 m AHD). Within the brines the point-water head is everywhere at 41.3 m, while the environmental-water head contours are vertical, showing no vertical gradient within the brine and therefore no movement. Despite the apparent horizontal gradient within the environmental-water head there is no water movement, since environmental-water head only determines whether vertical movement may occur. Similarly, the fresh-water head contours in the brine are horizontal, signifying no horizontal movement within the brine. The vertical fresh-water head 'gradient' reflects pressure levels in the brine insofar as they may be balanced by a column of fresh water; however, since fresh-water head cannot be used to measure vertical gradient within the brine, it does not indicate vertical flow.

12.9.1 Gradients and Velocities in Groundwaters of Variable Density

Given that the principal velocity components, v_x and v_y (horizontal) and v_z (vertical), coincide with the axes of the principal directional hydraulic conductivities, then:

$$v_x = -K_x [\partial H_{if} / \partial x] \quad (12.27)$$

$$v_y = -K_y [\partial H_{if} / \partial y] \quad (12.28)$$

$$v_z = -K_z [\partial H_{in} / \partial z] \quad (12.29)$$

Where K_x , K_y and K_z are the hydraulic conductivities in x, y and z directions, and ∂ denotes a differential. These equations permit the two horizontal flow velocities and the vertical flow velocity to be calculated. It is notable that whereas the hydraulic gradient, as defined by fresh-water heads, is used for the horizontal velocities, the environmental-water heads are used when calculating the vertical velocity. The actual flow direction is determined by the resultant of the three velocities. The value of environmental-water head in determining the vertical flow component is seen in the case of the Bourka 2/3 piezometer nest at southwestern Lake Tyrrell, where the point-water head in the deeper bore, Bourka 2, is 0.2 m less than that in Bourka 3, suggesting a downward gradient. However, the calculated environmental-water head difference is 0.5 m in the opposite direction.

An additional feature of this calculation, not normally occurring in other similar calculations in the Tyrrell Basin, is that since the piezometer nest is on the lake edge, the lake floor is the top of the zone of saturation from which point average density measurements are calculated. By re-defining the lake floor level to be zero datum, then equation 12.26 may be used as a simplified form of equation 12.24.

Calculation of environmental-water heads at the Bourka 2/3 piezometer nest.

TABLE 12.5 - Environmental water heads at the Bourka 2/3 piezometer nest

Bourka 2 bore

$$\begin{aligned}\rho_i &= 1.19, & Z_r &= 0 \text{ (where the lake floor is zero datum)} \\ \rho_a &= 1.175, & Z_i &= -62.3 \text{ m,} \\ \rho_f &= 1.03, & H_{ip} &= -0.24 \text{ m} \\ \text{Since } Z_r &= 0, & & \text{by equation 12.26 above} \\ \rho_f H_{in} &= \rho_i H_{ip} - Z_i (\rho_i - \rho_a) \\ H_{in} &= +0.63 \text{ m above lake floor}\end{aligned}$$

Bourka 3 bore

$$\begin{aligned}\rho_i &= 1.18, & Z_r &= 0, \\ \rho_a &= 1.17, & Z_i &= -17.3 \text{ m,} \\ \rho_f &= 1.03, & H_{ip} &= -0.04 \text{ m}\end{aligned}$$

here, $H_{in} = +0.12$ m above lake floor, that is, 0.51 m less than for Bourka 2.

(For $\rho_f = 1.00$, respective H_{in} values of +0.65 m for Bourka 2 and +0.13 m for Bourka 3 are obtained, giving a difference of 0.52 m.)

12.9.2 Luszczynski's Equation for the Salt Water - Fresh Water Inter-face

While flow directions and gradients are essential information in any understanding of the groundwater flow characteristics in the vicinity of Lake Tyrrell, a knowledge of the position of the salt water – fresh water interface within the transition zone is also useful in studying the lake - groundwater interactions. This can be obtained from detailed chemical data through the stratigraphic section; however, this information is not normally available. The Ghyben-Herzberg approximation assumes static conditions within the salt water, which may introduce considerable error.

Hubbert's equation 12.12 uses data from two points, one fresh and one salt, from some distance apart in a vertical column. However, it ignores any vertical flow component and is valid only where vertical flow is negligible and the interface is sharp. Luszczynski's equation for the theoretical position of the interface is derived from a more general equation which expresses the relationship between the point-water head in fresh water (therefore a fresh-water head) and the elevation of the contact between fresh water and diffuse water in a vertical sequence within an aquifer. Proof of the validity of the general equation is given in Luszczynski 1961, p. 4255). The interface equation is:

$$H_{1p} = \rho_i h + \rho_2 H_{2p} - Z_2(\rho_2 - \rho_a) - Z_d(\rho_a - \rho_1) \quad (12.30)$$

where h is the environmental-water head loss ($H_{1n} - H_{2n}$) between any two points -one in fresh water (1) and the other in salt water (2), in a vertical sequence in the aquifer containing fresh water, diffuse water (water in the transition zone) and salt water
 Z_d is the level of the contact between fresh water and diffuse water (transitional water). It is also the reference point from which p_a is measured
 Z_2 is the elevation of point 2 (m, AHD)
 H_{2p} is the point-water head at point 2 (m AHD)
 H_1 is the point-water head at point 1 (equals the fresh-water head—m AHD)
 ρ_i is fresh water density (g/cm^3)
 ρ_2 is salt water density (g/cm^3)
 ρ_a is the average density between Z_d and point 2 (g/cm^3)

The equation expresses a relationship between H_{1p} (a point-water or fresh-water head in fresh

water) and Z_d (the elevation between the contact between fresh water and the top of the transition zone). The first term on the right-hand side of the equation accounts for vertical flow in the section (as determined by environmental-water heads), the second term expresses the point-water head in the salt water, and the third and fourth terms account for variable density in the transition zone.

It follows that in situations where there is no vertical flow (that is $h = 0$) and where the interface is sharp with no transition zone (that is $p_a = p_2$) the third term vanishes, leaving:

$$\rho_1 H_{1p} = p_2 H_{2p} - Z_d(\rho_2 - \rho_1)$$

This equation is the same as Hubbert's equation for the interface. In a similar manner, Luszczynski shows that the Ghyben-Herzberg approximation is a further simplification of the general equation in which the point-water head in salt water is at sea level (or at lake level if a lake datum is used). Thus, $H_{2p} = 0$, and the Ghyben-Herzberg equation is obtained.

In order to accommodate the strong vertical flow component present in the groundwater discharge areas adjacent to the salt lakes, it is necessary to include the term $p_1 h$ in Hubbert's equation, thus giving:

$$\rho_1 H_{ip} = \rho_1 h + \rho_2 H_{2p} - Z_d(\rho_2 - \rho_1) \quad (12.31)$$

where Z_d is the elevation of the theoretical contact between salt water and fresh water.
 H_{1p} is the point-water head in fresh water, and therefore equals the fresh-water head ($H_{i.}$) of equation 12.13.

12.10 Applicability of the Australian Height Datum (AHD) and the Lake Level Datum in the Interface Equations

For inland settings, the application of the Ghyben-Herzberg equation to determine the depth of the salt water - fresh water interface beneath a lake system requires that a local datum be used as zero datum to calculate the depth to the interface. The datum used in this case is normally mean annual lake level. However, in salina situations, where the lake holds only a small amount of water for a minor part of the year, the level of the lake floor becomes the datum (it is assumed that the water table is close to the lake floor during the dry period). The choice of a local datum is not necessary for the Hubbert and Luszczynski equations, which maybe used whatever point is chosen as the datum from which all other elevations are measured. This allows all elevations in the calculations where these equations are used to be made relative to either the AHD or to mean lake level.

12.11 Characteristics of Groundwater Flow in the Timboram and Wahpool Sub-Basins

The Wahpool and Timboram lake basins, lying to the east of Lake Tyrrell, are considered for convenience as being discrete groundwater sub-basins, although the virtual absence of an intervening groundwater divide suggests that they are probably parts of a single groundwater sub-basin, receiving regional groundwater flow from the east and bounded on the west by the groundwater divide developed between Lake Wahpool and Lake Tyrrell.

The floor of Lake Timboram is at an elevation of about 46.5 m AHD, while that of Lake Wahpool is some 1.5 m lower at 45.0 m. One important difference between the two lakes is that Lake Timboram is the terminal lake to the Lalbert Creek, which like the Tyrrell Creek flows about once every decade. At these times the lake may hold 3-4 m of water and be sufficiently fresh to maintain a freshwater fish population for several years (Macumber, 1980). Lake Wahpool, which receives only direct precipitation and local runoff, contains less than 0.5 m of water at these times. The increased hydraulic gradients developed during the high lake regimes at Lake Timboram greatly facilitate groundwater flow between the two lakes and between the two sub-basins and the Tyrrell sub-basin.

Along the eastern shoreline of Lake Timboram, regional groundwater flowing from the east encounters significantly denser reflux brines infilling the aquifer beneath the lake and discharges into the lake basin in a similar manner to that occurring in coastal aquifers where a Ghyben-Herzberg interface exists. This process is repeated wherever regional groundwater flow enters either lake. Without outseepage through the lake floor, it can be readily calculated that, at the present rate of salt inflow into the Timboram-Wahpool system, over 5 m of salt would have accumulated on the floors of the two lakes over the past 16,000 years (that is, since the end of the last major phase of aeolian deflation). However, the virtual absence of any significant salt accumulation in these lakes indicates that reflux processes have also been operating. This is confirmed by the presence of the Timboram-Wahpool reflux brine, with a salinity range from about 95,000 mg/l to 125,000 mg/l, occupying the aquifer in the region between Lake Timboram and the eastern shoreline of Lake Tyrrell.

Lake brines, which pass downward into the Parilla Sand aquifer, flow westward towards Lake Wahpool, where some discharge into the lake occurs. Although no springs have been sampled on the eastern shoreline of Lake Wahpool, typical reflux brines occur in shallow piezometers (SR 70 and SR 77) situated on the southeastern shoreline of Wahpool, about 2 km west of Lake Timboram. The waters have salinities of 123,600 mg/l and 105,700 mg/l respectively (Table 12.6), and their point-water heads are 0.8 m and 1.0 m above the level of Lake Wahpool.

TABLE 12.6 - Composition of groundwater from the SR 70 and SR 77 piezometers and Moortworra 10001 bore (mg/l)

	Salinity	Cl ⁻	HCO ₃ ⁻	SO ₄ ²⁻	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	Fe ²⁺	Br
SR 70	123,600	69,350	184	8,950	1,248	5,034	38,390	240	2	174
SR 77	105,600	58,750	112	8,220	1,626	4,000	32,390	370	2	166
Moortworra10001	120,800	67,800	111	9,100	865	4,780	37,500	355	1	190
Average Timboram-Wahpool reflux brines	114,300	64,600	-	8,023	815	4,735	35,400	333	-	176

12.11.1 The Re-emergence of the Timboram-Wahpool Reflux Brines at Lake Tyrrell

While there is some discharge of Timboram reflux brines into Lake Wahpool, the bulk of the reflux brines pass westward to underflow the less dense regional groundwaters forming the broad groundwater divide between the Wahpool and Tyrrell sub-basins. Westward of the Wahpool groundwater divide there are no deep bores passing into the Timboram–Wahpool reflux brines until the edge of Lake Tyrrell is reached. It is surmised that there is a zone of mixing between the reflux brines and the regional waters as the interface rises in response to the decline in hydrostatic head in the regional groundwaters as Lake Tyrrell is approached. This would be analogous to upconing in coastal aquifers where over-pumping takes place. It is notable that, while Timboram–Wahpool reflux brines outflow along a broad strip of the Tyrrell spring zone on the eastern shoreline of the lake, there is no observed outflow of regional groundwaters in this zone. The latter waters are either entirely incorporated into the reflux brines or are lost from the system in the broad zone of halophytic phreatophytes which have developed on the eastern margins of Lake Tyrrell.

On reaching the eastern shoreline of Lake Tyrrell, the Timboram–Wahpool reflux brines, of S.G. 1.09, come into contact with the Tyrrell reflux brines, of S.G. 1.16, and are forced to the surface along a Ghyben–Herzberg interface (Figure 9.18). If the regional groundwaters had still survived as a discrete groundwater unit in the vicinity of the eastern Lake Tyrrell shoreline, near where the reflux brines outflow, it would require that two separate interfaces existed in the flow system at this point - one between the two reflux brines, and one between the Timboram–Wahpool reflux brine and the regional groundwaters. As in the case of the reflux brine and regional groundwater on the western shoreline of Lake Tyrrell, the interface between the two reflux brines is masked by a broad transitional zone. This is well shown in the SR 84 piezometer nest situated on the eastern shoreline close to the lake edge (Figure 9.18). It is notable that no regional groundwater occurs in this section, with the first water obtained (from 1.5 m depth) being the Timboram–Wahpool reflux brines. An indication of the character of the flow regime along the eastern shoreline of Lake Tyrrell is provided in the following table (12.7), which shows the vertical (upward) gradient component in the SR 84 piezometer nest, drawn from environmental-water head calculations.

TABLE 12.7 - Vertical hydraulic gradients in the SR 84 piezometer nest on the eastern shoreline of Lake Tyrrell

Piezo-meter	Z	H _{ip}	H _{ip}	ρ _i	ρ _r	ρ _a	H _{in}	H _{in}	∂H _n /∂Z	∂H' _{in} /∂Z
84A	40.8	43.07	42.72	1.07	1.03	1.07	43.07	42.72	+0.02	+0.26
84	38.3	43.06	43.24	1.105	1.03	1.09	43.13	43.34	+0.05	+0.04
84B	23.0	43.75	43.75	1.117	1.03	1.11	43.94	43.97		

H = point-water head in October 1981

H' = point-water head in April 1981

The gradient components are for intervals between the three bores.

The table shows that a fairly uniform upward gradient exists within the transition zone. However, close to the surface there are marked changes in the magnitude of the gradient (but not the direction). Apart from the high value of 0.26 m/m for October 1981, which reflects seasonal water table fluctuations, the remaining values are all within the range of values occurring on the western shoreline of the lake.

12.12 Characteristics of Groundwater Flow in the Vicinity of Lake Tyrrell

12.12.1 Groundwater Flow between the Parilla Sand and Deeper Aquifers

In order to determine the extent (if any) to which the deeper aquifer systems beneath the Parilla Sand were contributing towards groundwater discharge at Lake Tyrrell, a piezometer was placed into a more permeable zone within the Geera Clay, consisting of a thin shelly marl bed. The marl was encountered in the Bourka 1 bore at a depth of 150-170 m. The piezometer was artesian, and the water rose to a level about 1.5 m above the adjacent ground surface, or 3.7 m above the lake floor. The salinity of the waters was 78,000 mg/l, which is well below the salinity of the overlying Tyrrell reflux brines occurring in the Parilla Sand. Since vertical flow is to be measured, it is necessary to calculate environmental-water heads in order to obtain a value for the vertical hydraulic gradient. In this instance, where more saline waters overlies fresher waters, the environmental-water head value of 33.6 m (AHD) is less than the point-water head of 45.0 m (AHD).

Calculation of environmental-water head by equation 12.24:

$$\rho_f H_{in} = \rho_i H_{ip} - Z_i(\rho_i - \rho_a) - Z_r(\rho_a - \rho_f)$$

where $H_{ip} = 45.0$ m (AHD);

$Z_i = 42.0$ m;

$Z_r = -115.5$ m;

$\rho_f = 1.03$ g/cm³;

$\rho_i = 1.055$ g/cm³;

$\rho_a = 1.13$ g/cm³

i.e. $H = 33.6$ m (AHD), or 7.7 m below the lake floor level. (By comparison, if ρ_f is taken as 1.00 g/cm³, the value of H_{in} is 33.9 m AHD.)

For comparison, the fresh water head calculated by equation 12.22 is 48.9 m. The value of 33.6 m for the environmental-water head in the Bourka 1 bore may be compared with the value of 42.4 m in the Bourka 5 bore, situated about 10 m from Bourka 1. The Bourka 5 bore monitors, on interval, at about 20-25 m (AHD) in the Parilla Sand. Despite the apparent upward-directed gradient suggested by the point-water head, the environmental-water heads show instead that a downward-directed gradient exists. The environmental-water head difference of 8.8 m between the bores over the vertical distance of 130 m gives a downward gradient ($\partial H_{in}/\partial Z_i$) of 0.064 (using $\partial_f = 1.00$, $dH_{in}/dZ_i = 0.069$).

The vertical hydraulic conductivity of the Geera Clay is taken as 1.75×10^{-5} m/d, based on Lawrence (1975).

From equation 12.29, the downward vertical velocity (v_z) is:

$$v_z = K_z (\partial H_{in}/\partial Z_i) = 7.3 \times 10^{-7} \text{ m/d}$$

At these velocities the Geera Clay may be regarded as virtually impermeable for the purposes of this study.

12.12.2 Salinity Distribution beneath the Western Shoreline of Lake Tyrrell

The distribution of groundwater salinities beneath the western shoreline of Lake Tyrrell is shown in Figure 12.13. The section line commences at the lake edge in the south and runs along the shoreline for about 8 km before passing 1 km inland at the SR 80 piezometer nest. It continues northward, parallel to the shoreline, until the Bimbourie 4/5 nest is reached and then turns east back to the lake (Figure 12.17). On salinity distribution the diagram shows a threefold division of

the groundwaters into an uppermost regional groundwater regime, where the salinity is about 40,000 mg/l; an intermediate transition zone, with waters ranging in salinity from 40,000 mg/l to 240,000 mg/l; and a lowermost zone of undiluted reflux brines, with salinities ranging from 240,000 mg/l to 290,000 mg/l. By comparison with the other zones, the uppermost zone of regional groundwater is little more than a shallow lens above the transitional waters (Figure 12.18).

The shallowness of the zone of uncontaminated regional groundwaters partly reflects the broad transitional zone and the proximity of the lake shoreline. It also stems from the large density difference between the reflux brines and the regional groundwaters. As previously shown, the depth to the interface, using the Ghyben–Herzberg approximation, is only 7.9 m for each metre the water table rises above mean lake level. In Figure 12.18 the highest point above the lake to which the water table rises is about 2.7 m, giving an approximate depth to the interface, at that point, of only about 21 m. By comparison, in a coastal situation, where the fresh water density is 1.00 and the salt water density is 1.025, the depth to the interface, with a 2.7 m high water table, would be 108 m. The elevations of the interfaces may be calculated from the data given in Table 12.8. For instance, using the data for piezometer SR 78 and equation 12.30, the calculated depth of the interface is 38.7 m AHD, or 2.6 m below the lake floor. This value is supported by chemical data from the adjacent SR 78A piezometer, which monitors the interval 0.4 m higher than that calculated for the interface. The salinity at SR 78A is 144,000 mg/l, which is close to the estimated value of 150,000 mg/l at the interface.

The above examples show that on approaching the lake edge there is only a very thin layer of regional groundwater overlying the transition zone waters. On the marginal flats, where water tables are only slightly above the lake floor, the regional waters are virtually absent, and the transition zone waters are found at the water table. This pattern of salinity distribution is commonly observed around the lake edge, where groundwaters with salinities similar to those of the regional groundwaters are only found at the springhead, close to the inland limits of the marginal flats. On passing across the spring zone there is a rapid rise in salinity of the outflowing spring waters in response to the horizontal outcrop of the transition zone.

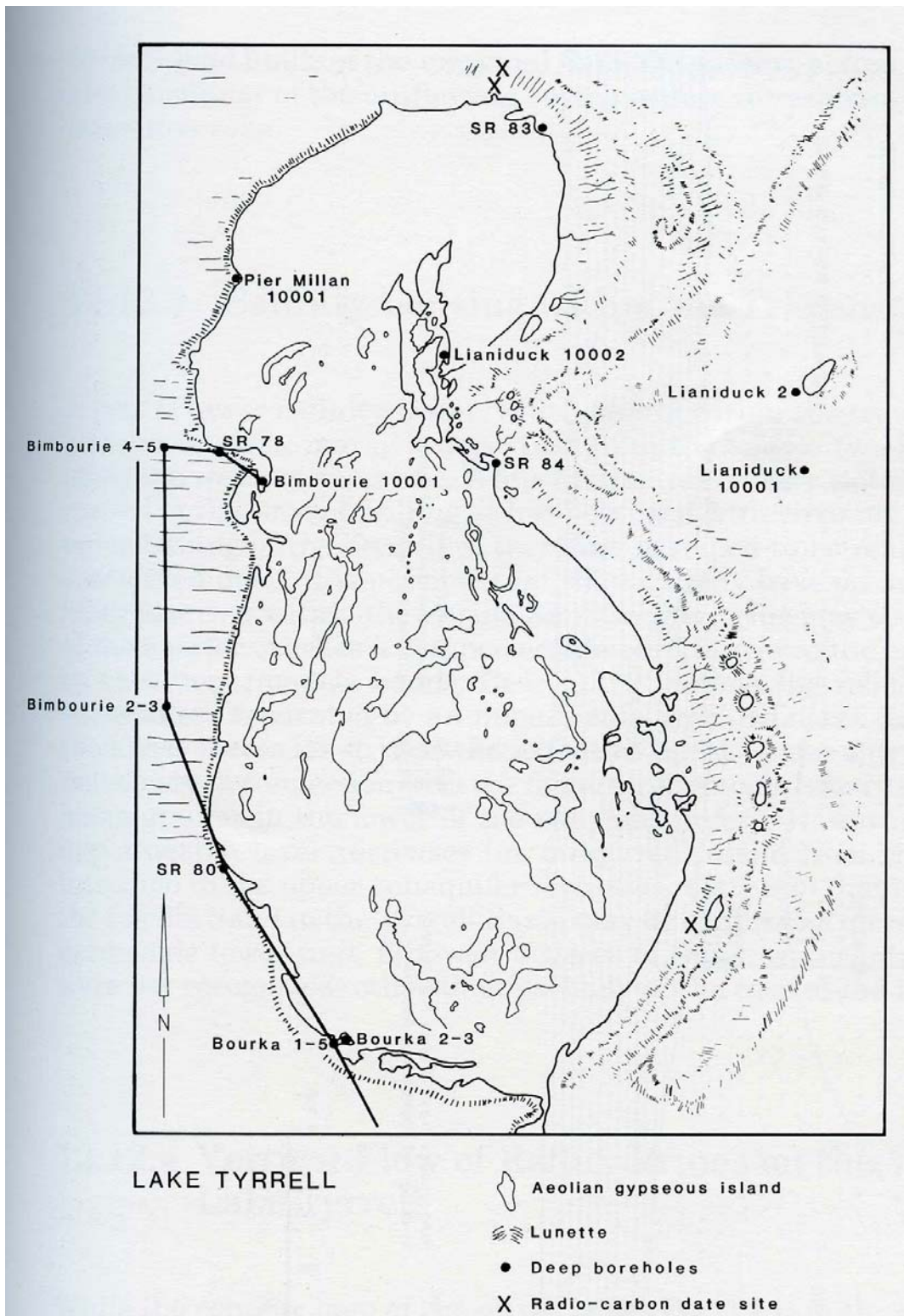


Figure 12.17 - Position of section line for Figures 12.13 and 12.18

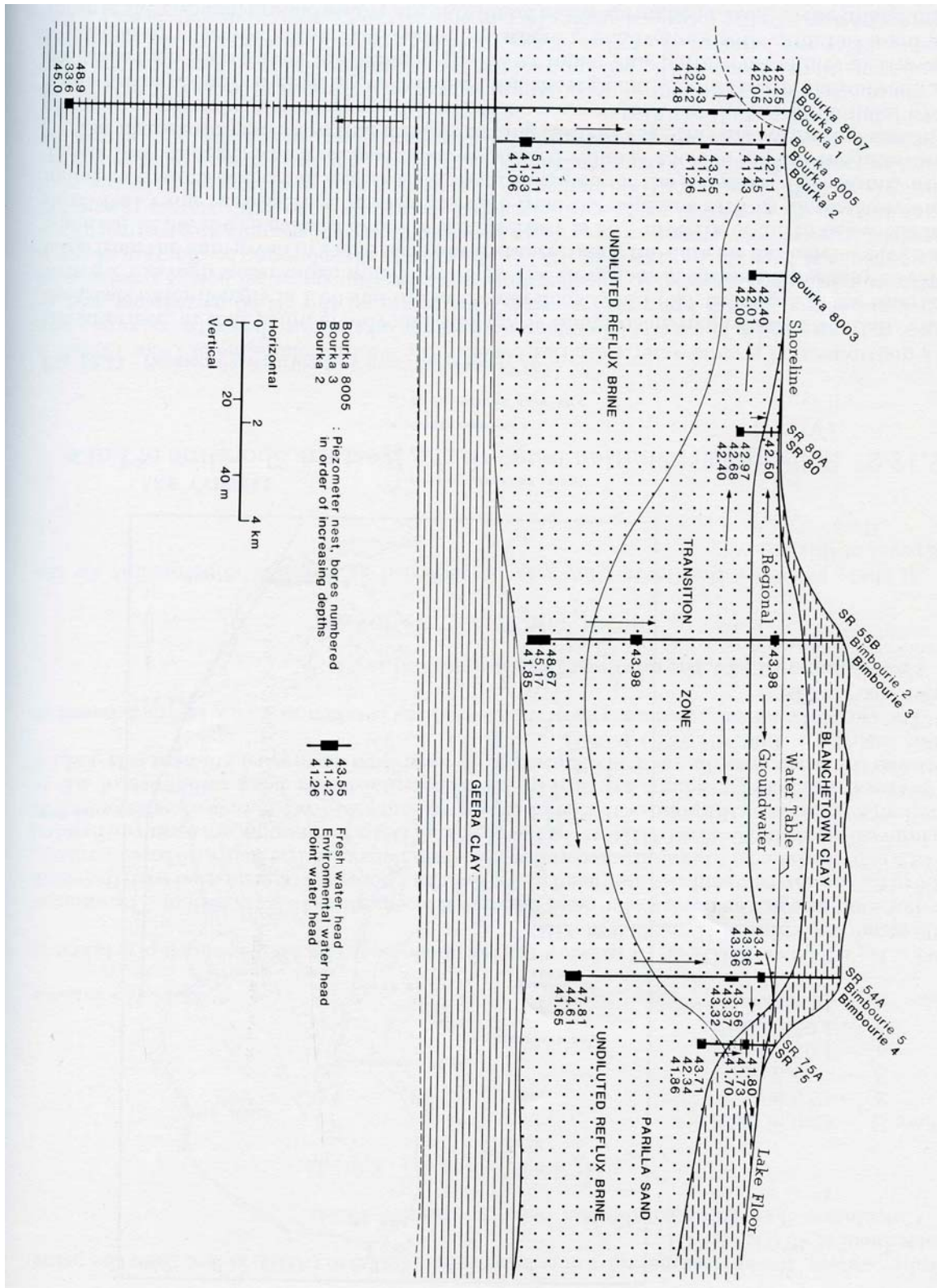


Figure 12.18 – Potentiometric head measurements and flow directions on the western shoreline of Lake Tyrrell (for section line see Figure 12.17).

to the inland limits of the marginal flats. On passing across the spring zone there is a rapid rise in salinity of the outflowing spring waters in response to the horizontal outcrop of the transition zone.

12.12.3 Salinity Lensing within the Transition Zone

A lens of lower salinity water, which occurs within the transition zone at the Bimbourie 2/3 piezometer nest, marks a departure from the simple two-layer system observed elsewhere along the western shoreline. While at first this fresher water was attributed to contamination caused at the time of drilling, it has persisted with virtually no change in salinity despite vigorous bailing of the bore. It is therefore assumed to be real. Although its origin can only be speculated upon, it is possible that the lens may have occurred as a consequence of permeability layering within the Parilla Sand aquifer. This may occur where a more permeable part of the aquifer overlies a less permeable part, or where the aquifer is divided into subaquifers by thin impermeable bands. Bear (1979) details the relationships that exist between two subaquifers separated by an impervious layer. He notes that as the total saturated aquifer thickness increases so does the extent of the seawater intrusion (an observation previously noted here in connection with the factors influencing lake closure to outseepage), the intrusion being greater in the lower of the two subaquifers. However, as the horizontal extent of the impermeable layer increases (as measured inland from the shoreline) then the degree of intrusion in the upper subaquifer increases relative to that in the lower subaquifer. Although the Parilla Sand in the Tyrrell Basin may be split into a more permeable upper unit and a less permeable lower unit, different degrees of saline water intrusion associated with each unit were not recognized, other than perhaps in the case of the Bimbourie 2/3 nest.

12.12.4 Vertical Flow of Reflux Brines on the Western Shoreline of Lake Tyrrell

While the contour map of the water table (Figure 12.1) in the vicinity of Lake Tyrrell indicates groundwater discharge from the regional groundwater system into the lake, an upward flow of deeper groundwater reflux brines is indicated by the height of point-water heads in the undiluted Tyrrell reflux brine, being commonly above the level of the lake floor. There is, therefore, at least a partial return flow into the lake of brines which have previously passed from the lake into the aquifer. The potentiometric head data for piezometer nests along the western side of Lake Tyrrell is shown in Figure 12.18. While the section provides a picture of the flow pattern, it should be borne in mind that it only represents a two-dimensional slice through the flow system. Since the Suction plane is approximately normal to the main downbasin flow direction, the horizontal flow components shown on the figure constitute only a minor part of the actual horizontal flow. Very little emphasis is therefore placed on the significance of these horizontal flow components. A more general picture is obtained from Table 12.8, which gives hydraulic head data and groundwater density data for selected bores in the Tyrrell Basin. The bores were chosen, firstly, by whether they constituted part of a piezometer nest, and, secondly, individual bores were selected where the groundwater densities were not those of the regional groundwater, and hence do not require density corrections.

TABLE 12.8 - Hydrostatic head and density data from selected bores in the Tyrrell Basin (for locations, see Figure 12.20).

Parish and bore number	ρ_f	ρ_i	ρ_a	H_{ip}	Z_i	Z_r	H_{in}	H_{if}	H_{2in}
Bourka 1	1.03	1.055	1.13	45.00	-115.5	42.00	33.60	48.89	33.35
Bourka 2	1.03	1.190	1.175	41.06	- 21.0	41.30	41.93	50.70	41.75
Bourka 3	1.03	1.180	1.17	41.26	24.0	41.30	41.42	43.77	41.43
Bourka 5	1.03	1.155	1.095	41.50	25.0	42.00	42.42	43.50	42.44
Bourka 8001	1.03	1.160	1.16	41.40	36.9	41.30	41.41	41.96	41.42
Bourka 8002	1.03	1.127	1.10	41.70	34.0	41.30	41.95	42.43	41.95
Bourka 8005	1.03	1.160	1.16	41.43	36.1	41.30	41.45	42.11	41.45
Bourka 8007	1.03	1.065	1.05	42.05	37.0	42.00	42.12	42.22	42.13
SR 80A	1.03	1.030	1.030	42.50	40.6	42.00	42.50	42.50	42.52
SR 80	1.03	1.098	1.065	42.40	34.0	42.00	42.68	42.95	42.70
Bimbourie 1	1.03	1.047	1.045	44.50	17.0	44.50	44.55	44.95	44.56
Bimbourie 2	1.03	1.062	1.10	43.90	10.0	44.00	42.64	44.95	42.60
Bimbourie 3	1.03	1.162	1.095	41.86	-12.0	43.98	45.17	48.76	45.21
Bimbourie 4	1.03	1.170	1.10	41.65	-4.0	43.40	44.63	47.85	44.67
Bimbourie 5	1.03	1.062	1.055	43.32	35.7	43.40	43.37	43.56	43.37
SR 54A	1.03	1.050	1.05	43.36	40.8	43.40	43.36	43.41	43.35
SR 55B	1.03	1.031	1.03	43.98	40.4	44.00	44.00	44.00	44.00
SR 78A	1.03	1.095	1.08	41.70	39.3	41.80	41.73	41.85	41.73
SR 78	1.03	1.158	1.125	41.86	27.0	41.80	42.34	43.71	42.36
SR 6023	1.03	1.030	1.03	44.50	43.0	44.50	44.50	44.50	44.50
Lianiduck 2	1.03	1.070	1.05	47.80	1.0	47.50	48.71	49.62	48.75
SR 83	1.03	1.079	1.08	44.79	37.7	43.10	44.88	43.44	44.91
SR 83A	1.03	1.079	1.08	43.17	42.1	43.10	43.17	44.93	43.17
SR 83B	1.03	1.043	1.065	46.58	25.0	43.10	46.25	46.90	46.35
SR 84	1.03	1.105	1.09	43.06	38.3	43.07	43.13	43.41	43.13
SR 84A	1.03	1.070	1.07	43.07	40.8	43.07	43.07	43.16	43.07
SR 84B	1.03	1.117	1.11	43.75	23.0	43.07	43.94	45.50	43.97

ρ_f is the fresh water density (1.03 g/cm³)

ρ_i is the point-water density (g/cm³)

ρ_a is the average water density between Z_i and Z_r (g/cm³)

Z_i is the elevation of screened interval in bore (top)-(m AHD)

Z_r is the elevation of the point on the water table from which p is calculated (= elevation of water table)-(m AHD)

H_{ip} is the point-water head at Z .

H_{in} is the environmental - water head (m), calculated by equation 12.24

* H_{2in} is the value if ρ_f is chosen as 1.00 g/cm³

H_{if} is the fresh-water head (m), calculated by equation 12.22

Apart from supplying information on the groundwater flow system, the table brings out many of the relationships between density, point-water head, environmental-water head and fresh-water head. Perhaps the most informative feature is that, with the exception of the Bourka 1 piezometer which monitors pressures from deep within the Geera Clay sequence beneath the Parilla Sand, all shoreline bores have environmental-water heads greater than the level of the lake floor (41.3 m). Vertical hydraulic gradients for a number of piezometer nests situated on the western shoreline are given in Table 12.11, while gradients for the SR 84 nest on the eastern shoreline are given in Table 12.7. A further set of environmental-water head values, based on a fresh water density of 1.00 g/cm³ is given in the last column of Table 12.8. Comparison with the equivalent head values, based on the regional water density of 1.03 g/cm³, shows that with only two exceptions the respective head values are all within 4 cm of each other, often being the same. It follows that vertical gradients based on either fresh water density value (1.03 or 1.00) will be very

similar.

TABLE 12.9 - Variations in head and vertical gradients at the SR 78 piezometer nest*

Piezometer	SR78		SR78A		$\partial H_{in}/\partial Z$
	H_{ip}	H_{in}	H_{ip}	H_{in}	
06-11-75	42.14	42.66	41.08	42.13	0.043
16-02-76	42.08	42.59	41.86	41.90	0.054
07-02-77	41.89	42.33	41.68	41.71	0.055
22-08-77	41.97	42.47	41.82	41.86	0.050
13-02-78	41.90	42.39	41.70	41.73	0.054
21-08-78	41.95	42.44	41.96	41.01	0.035
28-02-79	41.87	42.35	41.82	41.86	0.040
13-08-79	41.86	42.34	41.75	41.78	0.046
18-02-80	41.80	42.27	41.68	41.71	0.046
11-08-80	41.82	42.34	41.76	41.79	0.045
April 81	41.82	42.3	41.65	41.68	0.050
October 81	41.92	42.42	41.85	41.89	0.043

*Data from Rural Water Commission

Average winter lake level taken as 41.6 m (AHD)

Summer lake level taken as 41.3 m (AHD) = lake dry

$\partial Z_i = 12.3$ m

(N.B. the use of $\rho_f = 1.00$ makes a difference of +0.001 to the value $\partial H / \partial Z$)

In each case, average values have been chosen for point-water heads; therefore, the values in Table 12.8 represent averages for hydraulic head parameters. Variations in point-water head from those given in the table will alter both the H_{in} and H_{if} values, and, consequently, both the vertical and horizontal velocity components. In order to provide an indication of the degree to which head variations occur in the bores and how they affect vertical gradients and therefore flow velocities, data from two piezometer nests are given in Tables 12.9 and 12.10. The two nests are from the western shoreline of Lake Tyrrell, the potentiometric data being supplied by the Rural Water Commission. Additional vertical flow components were obtained by the use of the data from Table 12.8. Each nest covers a different vertical interval in the flow regime-the SR 78 nest has one piezometer virtually at the fresh water-salt water interface (SR 78A), and a deeper piezometer (SR 78), a further 12.3 m down, taps an interval within the undiluted reflux brines. The SR 80 nest has the deeper piezometer at the interface, while the shallow piezometer (SR 80A) lies just below the water table and monitors water table fluctuations in the regional flow system. The data chosen for comparison were those which would best reflect the range of variability in water levels, covering both seasonal variations and larger climatic fluctuations such as the 1975 wet period and the subsequent 1976 drought.

In both shoreline piezometer nests the point-water heads and environmental-water heads are all above the average winter lake levels, indicating a general groundwater flow into the lake throughout the year from the regional groundwater (SR 80A), the Tyrrell reflux brine (SR 78), and from the transitional water at the interface (SR 78A and SR 80). During exceptionally wet periods (about once a decade, as in 1975 when the lake reached levels of 41.8-41.9 m AHD), point-water heads in the aquifer brines may be matched by the lake level. At these times there may be short periods during which vertical hydraulic gradients reverse. Vertical gradients were calculated for five principal piezometer nests (Table 12.11). With exception of the SR 80 nest, the deeper piezometer monitors the point-water head within the undiluted reflux brines; in the SR 80 nest the deeper piezometer covers an interval in the transition zone close to the salt water - fresh water interface. In all instances the gradients show a vertical flow component towards the surface from within the reflux brines. The data in Table 12.11 (as in Table 12.81 is limited by its reference to only one set of piezometer data, as can be seen by comparison with Tables 12.9 and 12.10; it therefore provides only a guide to certain of the hydraulic parameters. However, the variations in

these parameters are only in degree, not in direction, and the more general variations in upward flow of the reflux brine is shown in the analyses in Tables 12.9 and 12.10. The observations in Tables 12.9, 12.10 and 12.11 are supported by hydrochemical data, which also indicates inflow of reflux brines into the lake (Chapter 9).

TABLE 12.10 - Variations in heads and vertical gradients at the SR 80 piezometer nest*

Piezometer	SR78		SR78A		$\partial H_{in}/\partial Z$
	H_{ip}	H_{in}	H_{ip}	H_{in}	
06-11-75	43.25	43.56	42.84	42.84	0.109
13-12-76	43.06	43.37	42.38	42.38	0.150
08-07-77	43.01	43.37	42.52	42.52	0.129
21-08-78	43.13	43.43	42.78	42.78	0.099
25-02-80	42.42	42.69	42.32	42.32	0.056
03-03-80	42.38	42.65	42.31	42.31	0.050
27-10-80	42.52	43.08	42.38	42.38	0.106
04-11-80	42.35	42.62	42.22	42.22	0.015
17-11-80	42.35	42.89	42.48	42.48	0.062
01-12-80	42.34	42.94	42.38	42.38	0.071
April 81	42.35	42.62	42.22	42.22	0.061
October 81	42.38	42.66	42.56	42.56	0.015

where $a_i = 6.6$ m

*Data from Rural Water Commission

TABLE 12.11 - Vertical gradients in groundwaters along the western shoreline of Lake Tyrrell (data from Table 12.8)

Piezometer	Z_i	H_{in}	$\partial H_{in}/\partial Z$	Salinity zone	Aquifer
SR 78A	39.0	41.73	0.05	Transitional	Blanchetown Clay
SR 78A	27.0	42.34		Undiluted reflux	Parilla Sand
SR 80A	40.6	42.31	0.05	Regional g/water	Blanchetown Clay
SR 80	34.0	42.65		Transitional	Parilla Sand
Bimbourie 5	36.0	43.37	0.03	Transitional	Parilla Sand
Bimbourie 4	-4.0	44.51		Undiluted reflux	Parilla Sand
Bourka 3	24.0	41.42	0.01	Undiluted reflux	Parilla Sand
Bourka 2	-21.0	41.93		Undiluted reflux	Parilla Sand
Bourka 8007	37.0	42.12	0.025	Transitional	Parilla Sand
Bourka 5	25.0	42.42		Undiluted reflux	Parilla Sand

12.13 Interactions Between Lake and Groundwater Systems

12.13.1 Introduction

While environmental-water heads at present show that there is a general upward movement of groundwater brines into Lake Tyrrell, this is a reversal of the hydraulic gradients which saw the downward migration of lake brines into the aquifer. During any reversal of gradients, an equilibrium point is reached where gradients are zero and hydrostatically driven flow ceases. Where there is no vertical flow in the aquifer, a vertical equipotential based on environmental-water head passes from the lake system to the base of the groundwater flow system (Figure 12.15).

12.13.2 Lake Outseepage Under Conditions of Uniform and Variable Density Waters

12.13.2.1 Nonsaline aquifer

Where the groundwater is uniformly fresh, outseepage occurs only if there is no continuity of the local flow system (Winter, 1976). In other words, outseepage requires that an equipotential extends from the lake surface to the base of the deeper groundwater flow system, whether it be an intermediate or regional system (Figure 12.19a and 12.19b). In the normal situation this is the base of the aquifer and the base of the regional flow system; however, where a number of lakes are involved, outflow from one lake basin to another may be via an intermediate flow system. Outflow occurs along a strip of varying width on the downbasin side of the equipotential.

In situations where the lake and groundwaters are both fresh, the vertical equipotential, which defines the threshold between lake outseepage and no outseepage, has a value equal to the elevation of the surface of the lake (Figure 12.19a).

12.13.2.2 Outseepage where a Ghyben–Herzberg interface exists

The requirements for outseepage are significantly different in situations where the underlying groundwater is sufficiently saline to have resulted in the development of Ghyben–Herzberg interfaces beneath the lake shoreline. Once established, these interfaces become flow boundaries to not only the local flow system but also to regional and intermediate flow systems. The brine body beneath the lake is essentially isolated from the adjacent fresh water flow system and has its own separate hydraulic characteristics, determined largely by its density, the aquifer parameters, and the depth and density of the water in the overlying lake system.

Since the brine is essentially a discrete water body, there is no continuity of an external groundwater flow path beneath the lake, and outseepage depends on the density and head relationships between the brine body and the adjacent regional fresh water system on the downbasin side of the lake. As the fresh-water head increases linearly with depth in brine of uniform density in the aquifer beneath a lake floor, lake outseepage commences whenever the fresh-water head at some depth beneath the lake floor exceeds the fresh-water head at a corresponding level beneath the enclosing groundwater divide. This most commonly occurs either seasonally, during winter, or during longer extended wet periods, when the lake fills, and thus further loads the brine column. It results in the extension downbasin of the saltwater wedge. Conditions conducive to basin outseepage occur if the fresh-water head at the base of the aquifer beneath the divide is exceeded by the fresh-water head at an equivalent level beneath the lake.

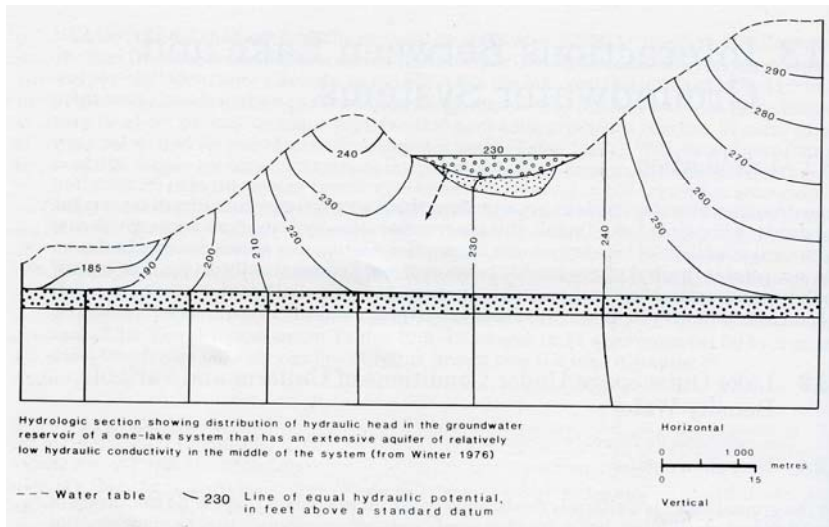


Figure 12.19a Diagram from Winter (1976) of a lake open to outseepage, showing a vertical equipotential extending from the lake to the base of the aquifer. There is therefore no continuity of the local flow system.

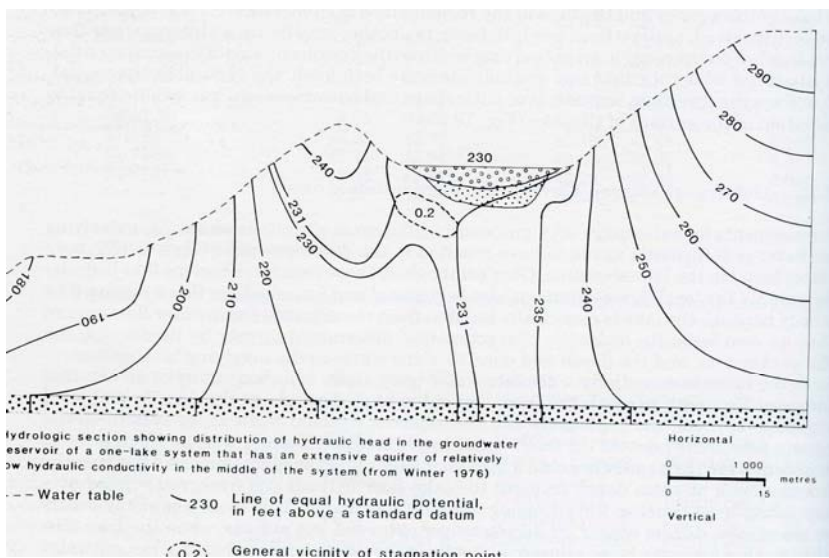


Figure 12.19b A lake closed to outseepage (from Winter, 1976) showing a stagnation point on the downbasin side of the lake. There is continuity of the local flow system beneath the lake and no vertical equipotential comparable to that in Figure 12.19a.

Where the closing fresh-water head is known, the point-water head beneath the lake, required to produce the equivalent fresh-water head, is readily found by equation 12.22. Given no vertical flow and a uniformly dense reflux brine, this point-water head defines both the threshold potential for outseepage and the corresponding lake level required to produce it. For instance, at Lake Tyrrell the closing fresh-water head produced by the groundwater divide is 45.55 m (Figure 12.6). The equivalent point-water head at depth beneath the lake, which is required to equalize this enclosing fresh-water head, is 41.28 m. In this example the same result is obtained by using equation 12.14, where the value of r_1 is 1.045 and r_2 is 1.16.

It is notable that this value is virtually the same as that of the elevation of the lake floor (41.3 m),

suggesting that gradients conducive to outseepage should come into being once water appears in the lake. This bears out the conclusion previously reached, using the Ghyben–Herzberg approximation, that if normal hydrostatic conditions prevailed in the aquifer the lake would be virtually on the threshold of outseepage even when it was dry, albeit with water tables at the level of the lake floor. However, in reality there is an upward flow in the reflux brines beneath the lake (as shown in Tables 12.9, 12.10 and 12.11), and the pressures at depth within the aquifer are sufficiently high to prevent lake outseepage under the present hydrological regime, except perhaps at times of highest lake levels. It is seen as a recent (albeit temporary) phenomenon and is discussed further in Chapter 12.14.

The development of the present situation, whereby the lake lies close to the threshold of outseepage, would have arisen gradually with the infilling of the aquifer by reflux brines under favourable hydrostatic heads. Given the presence of a groundwater divide, the necessary hydrostatic head requirements for lake outseepage could only come from additional water in the lake. For instance, with no reflux brines in the aquifer it requires about 3.7 m of lake water to enable outseepage with the present-day divide height. However, once the reflux brines are introduced into the aquifer, the head requirements (lake levels) simultaneously decrease until the present situation is reached.

An alternative hypothesis to the infilling of the aquifer by hydrostatic processes is that suggested by the work of Bachmat and Elrick (1970), who showed theoretically and experimentally that salt solutions may move from a reservoir into an underlying porous medium saturated with fresh water by a process of convective dispersion, arising from horizontal variations of salt concentration in the top surface of the porous medium, and to variations in the vertical velocity component. No attempt has been made here to assess the applicability of this process to salt lake systems, but if it were to occur to any extent, then the end result would be the same, that is, the eventual opening of the lake to hydrostatic head induced outseepage.

12.14 Abnormal Subsurface Pressures

Formation pressures which exceed hydrostatic pressure in an aquifer are termed surpressures (Fertl, 1976). Surpressures may arise from a number of causes including compaction, montmorillonite compaction, tectonics, hydrocarbon generation, artesian conditions, osmosis, palaeopressures and aquathermal effects (Magara, 1978). The development of abnormally high pressures requires at least some restriction to the migration of pore fluids; conversely, where hydraulically 'open' systems exist there is the development of hydrostatic conditions. Many of the possible causes of surpressures listed above are not applicable to the Tyrrell Basin, and of the remainder, two factors, osmosis and palaeopressures, are considered in this study.

12.14.1 Surpressures within the Tyrrell Reflux Brine

The general upward gradients occurring in the reflux brines show that heads within the aquifer beneath the lake and its close surroundings are greater than heads which can be produced by hydrostatic processes under the present-day hydrological regime. Furthermore, in certain instances, e.g. at the Bourka 2/3 piezometer nest where unusually high groundwater densities are found, it can be shown that the hydraulic heads are higher than those produced at the base of the aquifer beneath the groundwater divide. Indeed, to equalize the head produced by the reflux brine in the Bourka 2/3 sequence ($\rho_a = 1.175$ -from Table 12.8), it requires a groundwater divide height of 45.5 m (even where the regional groundwater density is taken as 1.045 g/cm^3 , as used in the example in the previous section).

As a consequence of the above pressure distribution within the aquifer, there is a slight pressure 'mound' developed at depth within the brines. It is suggested below that these anomalous pressures are subfossil or remanent pressures, established during phases of higher lake levels in the recent past, now undergoing decay, but reinforced at times of high lake level. Similar, albeit higher, surpressures occur within the Timboram—Wahpool brines.

The pressures present in the aquifer cannot be created or sustained by hydrostatic pressures generated under the present-day 'normal' hydrological regime in Lake Tyrrell. (The reverse is the case for surpressures in the Timboram-Wahpool brines, discussed later). In order to produce these pressures, it requires a slightly higher average lake level than occurs today. Some indication of the minimal requirement to produce these pressures can be obtained by examining the lake levels required to produce groundwater pressures similar to those now found in the aquifer. However, since the calculations show only what mean annual lake level is required to equilibrate with these aquifer pressures they provide, at best, only an insight into earlier long-term average lake levels.

Any lake levels suggested by the calculations must be minimal, since excess pressures found in the reflux brines would gradually decline with time. The presence of these pressures in the aquifer is interpreted as a further indication of a relatively recent change in the hydrologic regime, and more specifically in groundwater - surface water relationships caused by a small fall in average long-term lake levels. The fresh-water heads (H_f) and the depths (Z_i) of a number of piezometers tapping the reflux brine are given in Table 12.12. The environmental-water heads (H_{2n}) at an elevation beneath Lake Tyrrell corresponding to the fresh-water heads in the piezometers, give threshold values of the respective vertical equipotential, and, hence, the minimal lake level at the time that the higher pressures were established in the aquifer.

TABLE 12.12 - Hydraulic head data in piezometers to the west of Lake Tyrrell (m AHD)

Piezometer	Z _i	H _{if}	H _{2n}
Bourka 2	-21	50.7	41.9
SR 80	34	43.0	41.9
Bimbourie 3	-12	48.8	41.5
Bimbourie 4	-4	47.9	41.7
SR 78	27	43.7	41.8

Mean annual lake levels, ranging from between 41.5 to 41.9 m, are required to equilibrate with the pressures within the aquifer, thereby suggesting that mean annual lake levels in excess of 41.9 m existed to create these pressures by hydrostatic processes. The data from Table 12.12 shows that the surpressures beneath the lake are only marginally higher than those which could be obtained under the present-day regime, when peak winter lake levels are normally about 41.6 m AHD. Under the once-a-decade regime, abnormally wet periods cause lake levels to rise to a point where even the highest surpressures are equalled, if not just surpassed, and hence reinforced.

Given the marginal nature of the aquifer pressures over that of the present-day lake regime, the ability of the lake to equilibrate during the very wet years, and the evidence based on the thin salt crust of only a relatively recent closure to outseepage, it is considered that the premise, 'the pressures are subfossil pressures undergoing decline', is a valid assumption, albeit a tentative one. The marginally greater pressures within the brine provide an additional factor to that of the groundwater divide by further limiting outseepage from the lake. It is shown above (Table 12.12) that the excess pressures within the brines are equivalent to what could be produced under hydrostatic conditions, were the mean annual lake level about 41.5 m to 41.9 m AHD, that is, with an annual average of up to 0.6 m of water in the lake, or about 0.3 m higher than the present-day winter lake levels. A corollary of this conclusion is that lake levels in excess of that normally obtained are necessary before hydrostatically induced outseepage can commence at Lake Tyrrell under the present-day hydrological regime. It follows that these slightly higher brine pressures are perhaps the most important single factor controlling outseepage.

The above hypothesized higher average lake levels reflect pre-European conditions prior to the introduction of extensive land use changes. It is notable that these pressures are higher than those presently occurring at the groundwater divide; they therefore cannot have been created by the recent post-European rises in the water table documented earlier in this work. Given that the lake system has undergone a drying phase within the last 350 or so years (Chapter 12.3.4), it is possible that the slight pressures within the reflux brines are vestiges of higher pressures developed over a considerably longer time during the earlier lake-open regime, when, under higher lake levels, lake brines seeped into the aquifer to form the Tyrrell reflux brines. A similar effect, albeit on a very much larger scale, was described by Lloyd and Farag (1978), who noted the influence on hydraulic gradients of hydraulic head decay in arid regional drainage basins of North Africa and Saudi Arabia. Lloyd and Farag show that in certain instances fossil gradients reflecting the prior existence of ancient recharge mounds may persist for considerable time. In the case of the Tyrrell Basin, any tendency for a fossil head decay mechanism to be maintained for significant periods of time would be assisted by the presence of the Blanchetown Clay aquitard over much of the basin and by the occasional abnormally high lake levels during flood periods in the Tyrrell Creek.

12.15 Surpressures within the Timboram-Wahpool Reflux Brine

On the southern shoreline of Lake Wahpool, unusually high potentiometric heads are encountered deeper in the reflux brines; similar heads were found in reflux brines to the west of Lake Wahpool. For example, the point water level in the Moortwoora 10001 bore, at the southern tip of Lake Wahpool, is about 46.9 m (AHD), that is, 2 m above the floor of Lake Wahpool and 0.4 m above the floor of Lake Timboram. The bore taps an interval in the Parilla Sand, 15-27 m below the lake floor. As can be seen in Table 12.13 and Table 12.6, the waters are typical reflux brines, their density being 1.087 g/cm^3 . It is clear that these pressures are not being generated by hydrostatic processes emanating from Lake Wahpool and are therefore better equated with processes active upbasin at Lake Timboram. To some degree, the situation is similar to that occurring to the west of Lake Tyrrell, where high static levels in the Tyrrell reflux brines are considered to be reflecting a recent fall in lake levels. While this may also be true to some degree for the eastern basins, it is likely that the high static levels observed in the Timboram-Wahpool brines of the Wahpool sub-basin are partially maintained from pressures periodically generated upbasin at Lake Timboram during periods of high lake level. The transmission of hydrostatic pressures from the Timboram sub-basin to the Wahpool sub-basin is not inhibited by a Ghyben-Herzberg interface, as is the case for regional downbasin flow into Lake Timboram and Lake Tyrrell, or by an intervening groundwater divide. The Timboram-Wahpool reflux brines pass beneath regional groundwaters forming the broad groundwater divide to the west of Lake Wahpool, where they are encountered in two deep bores-Lianiduck 2 and Lianiduck 10001 (locality-Figure 12.17), having point-water heads of 47.8 m and 48.1 m respectively. In both examples, the static water levels are well above the level of Lake Wahpool (45.0 m) and are higher than the floor of Lake Timboram (46.5 m).

Using the method adopted at Lake Tyrrell (Chapter 12.14.1) to equate aquifer pressures at a given depth (Z_i) with the minimal lake level required to produce them, calculations were carried out on two bores—the Moortwoora 10001 bore on the edge of Lake Wahpool and the Lianiduck 1 bore on the groundwater divide to the west of Lake Wahpool—to establish what the equivalent Timboram lake levels might be. Data from each bore is given in Table 12.13, where the value H_{2n} is the minimal lake level which equates with the surpressures.

TABLE 12.13 - Potentiometric data from Lianiduck 2 and Moortwoora 10001 bores

	H_{ip}	H_{if}	Z_i	ρ_i	H_{2n}	ρ_{2p}	ρ_f
Lianiduck 2	47.8	49.6	1.0	1.07	47.1	1.087	1.03
Moortwoora 10001	46.9	48.3	21.1	1.087	46.9	1.087	1.03

H_{if} is obtained from equation 12.22;

H_{2n} is obtained from equation 12.23 where, $H_{2n} = Z_r$ = the lake level which equilibrates with H_{if} at point 'I' at elevation Z_i beneath the lake.

ρ_{2p} = the average density between point 'I' and the lake surface.

All hydraulic head values and depth values as m AHD.

The results obtained are 46.9 m in the case of the Moortwoora bore and 47.1 m for the Lianiduck bore, suggesting that the pressure levels in the reflux brines correspond to a lake elevation of about 47.0 m AHD. This is equivalent to an average annual depth of about 0.5 m of water in Lake Timboram. Therefore, the situation in the reflux brines to the west of Lake Wahpool is similar in some respects to that occurring at Lake Tyrrell, where small surpressures now determine the nature of the groundwater-lake interactions. One important difference is the significantly greater depth of water attainable in Lake Timboram than Lake Tyrrell following flow in the Lalbert Creek. For instance, lake levels at Lake Timboram may exceed 47.0 m AHD by 2 to 3 m; they were still about 47.6 m in August 1977, following the very high levels of 1973-75. The higher lake levels

occurring at Lake Timboram are more than sufficient to generate outflow pressures which override the surpressures and cause the lake to become a through-flushing system. This contrasts with the shallower Lake Tyrrell, which generally remains closed to outflow. The smaller size of Lake Timboram is the critical factor which determines periodic lake depth and therefore outflow potential. It is clear that, all else being equal, small lakes can more readily throughflush than large lakes like Tyrrell.

Beneath the Wahpool groundwater divide (formed by regional groundwater flow), the 0.5 m excess pressure in the reflux brines in turn creates an ephemeral second deeper groundwater divide within the reflux brines. During those times when low lake levels permit the required eastwardly directed gradients, brines flow both westward in the direction of Lake Tyrrell and eastward towards Lake Timboram. The explanation given for the surpressure in the brines beneath the Wahpool groundwater divide is the same as that proposed at Lake Tyrrell; that is, the surpressures are subfossil and reflect a relatively recent modification in the long-term, average surface water regime of Lake Timboram. However, in the case of Lake Timboram, the prolonged periods for which the lake may be held at levels above 47 m (AHD) implies that the dissipation of these surpressures may be slowed, or stopped, or even reversed during wet periods.

12.16 Osmosis - A Possible Alternative Cause of Subsurface Pressures

At various times during the course of the investigation attention was drawn to the potential for osmotic processes to be active within the groundwater systems. Although this arose out of consideration of the wide salinity range within the aquifer, it came into sharper focus as a result of the slightly higher pressures in the reflux brines, above those obtainable under present-day hydrostatic processes. An examination of the likely influences of osmotic processes within the Tyrrell Basin is discussed below. It is essential for any analysis of the flow system beneath Lake Tyrrell to determine whether the surpressures are osmotically induced. If they are, then the vertically upward hydraulic gradients found beneath the lake shoreline and elsewhere do not necessarily indicate flow directions, since osmotic processes create higher hydraulic pressures in the direction of the osmotically induced flow. Equilibrium is reached only when hydrostatic heads become equal to the osmotic driving forces. Furthermore, if the surpressures are of osmotic origin, they will remain constant once equilibrium between the osmotic and the hydrostatic forces has been reached; if they have not yet reached equilibrium, they will increase gradually with time. On the other hand, if the pressures are palaeopressures they are continuously undergoing decline. This will eventually lead to a re-opening of the lake to outseepage (at least seasonally) by normal hydrostatic forces and the re-emergence of the groundwater divide as the primary factor which determines the susceptibility of the lake to outseepage. It has long been accepted that shales or clays may serve as semi-permeable membranes for the purpose of osmosis, and this process has been variously invoked to explain anomalous formation pressures in many areas throughout the world, such as the San Joaquin Valley, California (Bailey et al., 1961), the Western Sedimentary Basin of Canada (Berry and Hanshaw, 1960; Hill et al., 1961; and Van Everdingen, 1968) and the Illinois Basin (Bredehoeft et al., 1963).

The osmotic pressure developed across such a clay membrane is roughly proportional to the salinity differential of the groundwaters present on either side of the membrane (Glasstone, 1946; Babcock, 1963; Jones, 1968). Hill et al. (1961) showed by experiment that osmotic pressures across a highly compacted shale may be of the order of 83 to 103 kPa (0.84 to 1.1 kg/cm²) for each 1000 ppm difference in water salinity. A similar relationship between osmotic pressure and water salinity was demonstrated by Jones (1969), who produced graphs showing the theoretical osmotic pressures to be expected across clay membranes. Jones (1968) noted that the pressure difference across a simple clay bed could, under natural conditions, exceed 3500 psi (246 kg/cm³). He comments that in known geopressed reservoirs, stepwise increments of osmotic pressure with depth through a series of bedded sands and clays could, as by a multistage pump, produce any of the observed reservoir pressures in the northern Gulf of Mexico. Babcock (1963), on the basis of thermodynamic considerations, suggested that the osmotically induced pressure differential across semi-permeable membranes could be deter-mined by

$$P_o = (RT/H_2O^{1/2}) (1n[H_2O_i/H_2O_{ii}]) \quad (12.32)$$

where

P_o is the hydrostatic pressure differential

R is the gas constant T is degrees Kelvin

H_2O is the molal volume of pure water

H_2O_i and H_2O_{ii} are the activities of water in the more saline solution and less saline solution respectively.

Freeze and Cherry (1979) show that, with the salinity concentration across a clay membrane varying from 12% NaCl to 6% NaCl, a pressure difference equivalent to 694 m of hydrostatic head is theoretically obtainable by equation 12.32. This salinity variation is not too dissimilar to that between the Timbora-Wahpool reflux brines and the regional groundwater.

12.16.1 Propensity in the Tyrrell Basin for Osmotic Mechanisms

While large salinity differences may exist within the Tyrrell Basin groundwaters, the Parilla Sand is the only significant aquifer, and apart from the Blanchetown Clay which acts as an aquitard over much of the basin, there is no development of any thick shale or clay sequence. At best, there are thin clay lenses, largely restricted to the lower less permeable zones of the Parilla Sand. One possible example of the influence of clay layering within the Parilla Sand is seen in the presence of lower salinity water sandwiched between brines in the Bimbourie 2/ 3 sequence, about 1 km west of Lake Tyrrell (Chapter 12.12.13 and Figure 12.13).

Vertically upward hydraulic gradients, indicative of higher pressures at depth, are of the same order of magnitude whether they are measured within the Parilla Sand beneath the Blanchetown Clay (Bimbourie 4/5 nest), or within the Parilla Sand where the Parilla Sand is unconfined (Bourka 2/3 nest), or where the piezometers span both the Parilla Sand and Blanchetown Clay (Table 12.11). Therefore, if osmotic processes are producing surpressures, their operation is not dependent on the presence of the Blanchetown Clay, but is equally due to clay lensing within the Parilla Sand.

Should surpressures have been produced within the Parilla Sand, the presence of an enclosing aquitard such as the Blanchetown Clay would be important for the retention of higher pressures. However, while for much of its length the Tyrrell Basin is confined by the Blanchetown Clay, it is not closed at its downbasin limits by any stratigraphic trap, and it is doubtful if the simple permeability decrease provided by the Tyrrell Fault would provide any real impediment to pressure release were this to be generated within the basin. Similarly, the absence of Blanchetown Clay from the southern areas of Lake Tyrrell provides a further zone where pressure release might take place. Indeed, although pressures within the brines are slightly higher than those which might be produced by the present-day hydrologic regimes by hydrostatic processes, they are not so excessive as to suggest that a closed system conducive to high surpressures exists within the Tyrrell Basin. That is, high surpressures would not be expected even if osmotic processes were operating. While the high salinity differentials of about 200,000 mg/l existing to the west of Lake Tyrrell have (under the right conditions) the capacity to generate very high osmotic pressures, the highest pressures reached are only about 0.7 m of hydrostatic head, or 7 kPa (1 psi) above the equivalent non-osmotic hydrostatic head. This suggests that either osmotic pressures are not being generated, or, if osmosis is proceeding, pressures are being rapidly dissipated on the saline side of membranes.

If pressure dissipation is occurring on the downflow side of thin clay membranes within the Parilla Sand, evidence for the existence of the osmotic processes might be found in abnormal chemical ratios within the aquifer, which are produced by selective filtration of ions by the clay membranes.

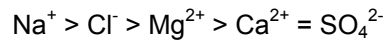
12.16.2 Ultrafiltration

De Sitter (1947) suggested that an osmotic mechanism could explain the presence of brines in groundwater basins where there is no obvious source of salts. The process of ion filtration by clay membranes is called ultrafiltration or salt sieving (Back and Hanshaw, 1965) - it has been previously referred to in Chapter 5.5.

The salt filtering properties of clay membranes were demonstrated by Buneev et al. (1947), Lumtadze (1954), Kemper (1960), McKelvey and Milne (1962) and various others. The chemical modification of the brines during migration across a membrane was discussed by Siever and Garrels (1961), Warner (1964), and Siever et al. (1965). White (1965) proposed specific mechanisms which could determine the rate of migration and the chemical differentiation across a clay membrane. A review of the relative factors influencing membrane filtration is given by Berry (1969).

The general tendency in certain sedimentary basins for the Ca/Cl ratio to increase with salinity

was interpreted by White (1965) as a function of selective ionic sieving through clay or shale membranes. In White's examples the process is reverse osmosis, where water is driven through the membrane by hydrostatic force. White defined the mobility of ions as their relative ability to pass through the clay membrane, noting that, in the case of Ca^{2+} and Na^+ , the larger divalent Ca^{2+} ions tend to be adsorbed on the cation exchanging clay, while the smaller monovalent Na^+ ions pass across the membrane. He proposed that the decreasing order of mobilities is:



Berry (1969) noted that in all geological environments where he deduced that membrane ultrafiltration exists, the Ca^{2+} ion was being ultrafiltered with respect to the Na^+ ion. Berry commented that clays are cation exchangers formed by singly charged SiO^- and AlOSi^- sites, and that some minor anion exchange occurs with replaceable OH^- ions. The clays adsorb divalent cations in preference to monovalent cations. The halogens substitute for OH^- in the clays.

12.16.3 Application to the Tyrrell Basin

Calcium ratios must be used with extreme caution if employed as indicators of membrane filtration in the Tyrrell Basin. This arises because of the presence of a wide range of Cl/Ca ratios in the groundwaters of the basin. For instance, precipitation of gypsum during the evaporative concentration of lake waters leads to high Cl/Ca ratios in the subsequent reflux brines; conversely, re-solution of evaporite calcite leached from the lunette systems leads to lower than normal Cl/Ca ratios in the underlying groundwaters. Apart from Na/Ca and Cl/Ca ratios, the other ratios most applicable to the waters of the Tyrrell Basin are Na/Mg , Cl/Mg , and Cl/Na .

One situation where ultrafiltration processes might be expected to occur is at the Bimbourie 2/3 piezometer nest to the west of Tyrrell, where a 'fresh' water lens of salinity 77,000 mg/l is sandwiched between brines of salinity 160,000 mg/l (above) and 250,000 mg/l (below). The vertical interval spanning these salinity changes is about 30 m (Figure 12.13). The presence of the 'fresh' water lens suggests the existence of zones of variable permeability within the aquifer at this point. Four water samples were taken at different times from the 'fresh' water, while a further four samples were taken from the underlying brine. The salinity range between the two sample points was about 170,000 mg/l over a distance of about 22 m. Comparisons of the ionic ratios between these two waters are shown in Table 12.14.

TABLE 12.14 - Ionic ratios of groundwaters from the Bimbourie 2/3 piezometer nest

	Bimbourie 2	Standard deviation	Bimbourie 3	Standard deviation
TDS (mg/l)	78,000		250,000	
Depth (m)	50		72	
Na/Ca	61.0	3.9	166.5	16.5
Na/Mg	8.8	0.2	8.1	0.2
Cl/Mg	14.4	0.5	14.0	0.4
Na/Cl	0.61	0.01	0.58	0.02
No. of samples	4		4	

Comments on Table 12.14

Should selective ion filtration occur as a consequence of osmotic processes, then decreased Na/Cl , Na/Mg , Na/Ca , and Cl/Mg ratios might be expected on the 'upstream', or 'fresh' water side of the membrane, while increased ratios would be found on the brine side. The table shows that significantly low Na/Ca ratios do occur on the freshwater side of the membrane. However, the Na/Mg and Cl/Mg ratios are marginally higher in the 'fresh' water, as is the Na/Cl ratio.

While there is some scope for interpretation of the Na/Ca ratios, the Cl/Na, Na/Mg, and Cl/ Mg ratios unambiguously suggest that there has been little or no selective ion filtration between the freshwater and saline formations. The lower Na/Ca ratio lies within the range normally found in the Tyrrell reflux brine, and may be adequately explained by the initial variations in calcium distribution in the transition waters. A similar approach was adopted to examine a number of piezometer nests where higher than normal pressures are found in the Tyrrell reflux brines. In each example, comparisons of ionic ratio are made across a vertical interval, between a deeper higher pressure zone and a shallower zone (Table 12.15). Three piezometer nests were chosen in order to cover the range of water salinities and lithologies in the basin. The nests, all situated on the western shoreline of Lake Tyrrell, are: SR 78 nest, at northwestern Lake Tyrrell; SR 80 nest and the Bourka 2/3 (4) nest, at southwestern Lake Tyrrell (for location, see Figure 12.17).

TABLE 12.15 - Ionic ratios for selected piezometer nests-Lake Tyrrell

Piezometer Stratigraphy	78 Par. Sand	78A Bl. Clay	80 Par. Sand	80A Bl. Clay	Bourka 2 Parilla	Bourka 4 Sand
TDS (mg/l)	235,000	144,000	150,000	43,000	290,000	265,000
H _{in} (m AHD)	42.3	41.7	42.6	42.3	41.9	41.4
Z _r (m AHD)	27.0	39.0	34.0	40.6	-21.0	24.0
Na/Ca	185	196	66	34	216	153
Na/Mg	9.7	9.1	11.8	15.4	9.7	9.8
Na/Cl	0.59	0.59	0.58	0.62	0.58	0.59

Comments:

SR 78 Nest-SR 78 contains reflux brines and lies within the Parilla Sand; SR 78A is situated near the top of the overlying Blanchetown Clay and lies in the transition zone virtually at the theoretical interface between the regional groundwater and the reflux brine. The Na/ Ca ratio is in the reverse order to what should occur if selective filtration is occurring under osmotic conditions-however, no significance is attached to the difference in ratios between the two piezometers, and the data suggests that ultrafiltration is not operating. The Na/Cl ratio also indicates no selective ion filtration. The Cl/Mg and Na/Mg ratios, although in the right order for selective ion filtration, are not sufficiently different to prove such filtration.

SR 80 Nest-SR80, the deeper piezometer, monitors an interval at the interface within the transition zone (it is in the Parilla Sand); SR 80A lies in the Blanchetown Clay close to the surface and contains the inflowing regional groundwater. Na/Ca ratio: this is in the right order for ultrafiltration; however, the ratio disparity is readily explained by the normal calcium distribution across the transition zone. The Cl/Na, Cl/Mg and Na/Mg ratios are all in the reverse order for selective ionic sieving, and therefore suggest that this is not occurring.

Bourka 2 /3 (4) Nest-both piezometers lie in the Parilla Sand, at southwestern Lake Tyrrell. The Parilla Sand outcrops in the lake floor and there is no confinement by Blanchetown Clay. The Bourka 4 bore has been chosen in preference to Bourka 3 bore since a more complete set of chemical data is available. Bourka 3 and Bourka 4 are at the same site and they both monitor the same stratigraphic interval, the former being used as an observation bore when carrying out

pumping tests on the Bourka 4 bore. The Bourka 2 piezometer taps an interval close to the base of the Parilla Sand and contains the most saline groundwater encountered in the Tyrrell Basin. Na/Ca ratio: significantly different values between the two piezometers may indicate selective ionic sieving; however, as in the above case, this may be equally well explained by initial Na/Ca difference in the reflux brines. The Na/Mg, Cl/Mg, and Cl/Na ratios are virtually the same in each piezometer, and therefore show that the respective ions are not being differentially sieved between the piezometers.

In conclusion, it is clear that despite the marked vertical salinity gradients in the above piezometer nests, there is no chemical evidence that selective membrane sieving occurs in the Tyrrell Basin.

12.16.4 Osmotic Processes in the Tyrrell Basin-Discussion

As is discussed above, there are many factors which can cause abnormal pressures within an aquifer; however, two things must be stressed-firstly, that the observed pressures are very small, and secondly, that they are only a recent innovation, preventing outseepage only for the time it has taken to accumulate a 70 mm salt crust. These factors point to subfossil pressures rather than the alternatives (including osmosis) for the closure of the lake to outseepage. Despite the high salinity differentials of groundwaters in the Tyrrell Basin, which would be expected to provide a strong potential for the development of both osmotically induced high pressures and the attendant chemical differentiations within the Parilla Sand aquifer, there is no hydrogeological evidence that either of these effects is present to any significant degree, if at all.

Taken alone, the absence of high aquifer pressures, commensurate with the high salinity differentials, may be explained by pressure dispersion on the saline side of any clay membrane present; however, when taken together with the hydrochemical data which contains no evidence of ultrafiltration effects, then it appears that osmotic mechanisms are not significant in influencing groundwater flow in the Tyrrell Basin. The absence of osmotic induced effects within the aquifer is attributed to the lack of suitable clay layers to form membranes. Where clay layers exist, as in the case of the Blanchetown Clay, it seems that their permeability is such as to render them unsuitable as membranes. In the case of clay lenses in the Parilla Sand, their discontinuous nature is apparently also against their use as an effective semi-permeable membrane.

12.17 Lake and Groundwater Interactions in the Tyrrell Basin Discussion

The groundwater divide in an arid or semi-arid region comes into being when groundwater discharge lakes behave as groundwater pumps operating on a sloping water table. As a consequence, the groundwater divide is relatively low, and may be set well back from the lake - in the case of Lake Tyrrell about 6 km, and for Lake Wahpool about 4 km. The absence of a comparable divide on Lake Timboram reflects the close proximity of Lake Wahpool (2 km further downbasin) and the long periods for which Lake Timboram may hold water.

In the case of a single lake, the groundwater divide, if present, is commonly independent of any topographic divide, which is usually restricted to the lake margin. The very low hydraulic gradients and small local recharge component permit the Dupuit-Forcheimer assumption (that the groundwater flow is essentially horizontal) to be used for flow at the divide. The validity of this usage is supported by potentiometric data. Therefore, the vertical equipotential having the same value as that of the water table elevation at the divide may be used to determine the closing head which must be overcome before outflow from the lake occurs.

Undoubtedly, in terms of this study, the most significant difference in the lake-groundwater interaction between lakes in temperate areas and those in arid or semi-arid regions is the process whereby in the latter systems, concentrated lake waters may recharge the groundwater systems beneath the lake whenever favourable hydrological conditions arise within the basin. This may come about by seasonal rises in lake levels. It may, in turn, lead to the gradual accumulation a high density brine body beneath the lake basin, as is the case with many large groundwater discharge lakes and boinkas of northwestern Victoria. In these cases the presence of a downbasin groundwater divide is not the significant factor in limiting outseepage from the lake system, as it is in the case of lakes in temperate areas of uniform groundwater density, such as that investigated by Winter (1976). In the case of the Wahpool groundwater divide, high pressure levels from upbasin, coupled with high brine density, allow ready underflow of the divide. The effect of the divide on interbasin flow is merely to depress the freshwater – saltwater interface, but since the base of the aquifer is at a much greater depth, outseepage is largely unhindered by its presence.

At Lake Tyrrell, the groundwater divide imposes a very delicate balance on the flow system within the Tyrrell sub-basin, whereby small changes in either lake level or water divide elevation affect the direction and the degree of the interaction between the lake and the aquifer. However, while the height of the divide and the salinity differential normally determine the threshold potential for outflow, other factors, such as the depth to the base of the aquifer beneath the groundwater divide, are also important considerations governing outflow. Yet even where all the above parameters are known, the situation may arise, as it has in the Tyrrell Basin, where relatively small vagaries in the lake hydrology require a readjustment in the hydrological equilibrium between the lake and the groundwater system. During this period, the interaction between the lake and the groundwater system is not solely determined by the above factors, but is modified by more intrinsic pressure-head relationships, developed within the aquifer in response to these changes in lake hydrology. This is a palaeohydrologic factor and relates to the previous history of the lake. In the case of Lake Timboram, the basic nature of the lake-groundwater interaction is not greatly affected; however, at Lake Tyrrell there has been a temporary reversal in the flow direction of the reflux brine, leading to the accumulation of the salt crust. Once the readjustment has taken place and the 'normal' hydrostatic regime prevails, the gradient will reverse and throughflow conditions will be established, at least seasonally. The salt crust will then gradually disappear from the lake floor.

13. CONCLUSIONS

The structural and stratigraphic framework of the regional aquifer system in northwestern Victoria was established in the Tertiary period and has undergone little change since that time.

Regional Groundwater Flow

At present, as in the past, regional groundwater flow passes northward, from recharge areas in the highlands toward regional discharge zones situated in the lower Loddon Plain and in the Mallee. The principal regional aquifers are the Lower Tertiary Warina Sand of the Renmark Group, the Late Tertiary fluviatile valley-fill gravels and sands of the Calivil Formation (now buried beneath 60 m to 80 m of Quaternary sediments) and its downbasin marine equivalent, the unconfined Parilla Sand.

Where groundwater discharges at the surface it affects the geomorphology, the soils, and the vegetation. For instance, in a riverine landscape, springfed effluent streams may commence flowing during times of higher pressures within the deeply buried, semi-confined Calivil Formation aquifer. These streams have a distinct geomorphic and sedimentologic character, which distinguishes them from other streams on the plain, and they are invariably brackish or saline. It is suggested that the presence of a zone of regional groundwater discharge across the path of a river system may strongly influence stream morphology. Further downbasin, in the Mallee region, groundwater discharge commonly occurs in small salinas and gypsum flats scattered throughout the linear dunefields, and in the larger groundwater discharge complexes - the boinkas. The Raak boinka is over 800 km² in area.

The groundwaters of northern Victoria are essentially a sodium chloride type, and, on passing downbasin, the chemistry rapidly evolves towards that of seawater. The ionic ratios of the groundwaters and surface waters suggest that there is an 'oceanic' overprint everywhere present from an early stage of their chemical evolution. Of the processes available, the most satisfactory explanation encompassing the high chlorinity and oceanic character is that of atmospheric salt accession, that is, cyclic salts. Whatever the salt source, it is clear that the evolutionary end point for Murray Basin groundwaters is one where the major ions Cl⁻, Na⁺, Mg²⁺, Br⁻ and to a lesser extent SO₄²⁻ attain relative ratios similar to that occurring in seawater.

The Riverine Plains

On the plains, regional groundwater discharge occurs in the Loddon Valley, at the junction of the continental suite with the marine sequences. An important factor in the development of a zone of regional groundwater discharge in the lower Loddon Plain is the change in hydraulic conductivity across the regional groundwater flow path, brought about by the replacement of highly permeable channel gravels and sands of the non-marine sequences by the significantly less permeable marine sands. A second factor contributing to regional groundwater discharge arises from the nature of the aggradation in the Loddon Valley. Infilling of the deep lead trench was in response to base level rises during the Late Tertiary marine transgression. This resulted in an upbasin thickening of the continental suite. The resulting downbasin wedging of the coarse clastic regional Calivil Formation aquifer leads to the development of a regional groundwater discharge zone in the lower and central Loddon Plain. As a consequence of these two geological controls, almost the entire central and lower Loddon Plain has artesian or near-artesian groundwater pressures which convert the plain into a broad zone of regional groundwater discharge.

Despite the aquifers on the plains being confined by 60 m of Quaternary sediments, there is nevertheless a ready interaction between the regional groundwaters and the surface systems, resulting in a delicately balanced hydrological equilibrium.

The delicate hydrological balance within the groundwater system was demonstrated by the rapid 20 km upbasin expansion in the discharge zone of the Loddon Valley in response to several abnormally wet years during 1973-75. In the adjacent Campaspe Valley there has been a uniform

rise of about 0.25 m/year in potentiometric head of the regional Calivil Formation aquifer since the turn of the century, and steady rises of from 0.1 to 0.2 m/year in potentiometric head in the regional aquifers of the Goulburn and Murray Valley systems. Within the Mallee region, rises in the water table associated with the unconfined Parilla Sand aquifer have caused the salinization of the lower depressions, with the reactivation of salinas and gypsum flats in areas which were formally grassed and had tree cover.

The rising water tables now being experienced across northern Victoria, and the subsequent salinization of large areas of both irrigated land and dryland, are seen as a result of increased accessions to the groundwater systems brought about by the removal of trees and the introduction of various agricultural practices, such as irrigation, following European settlement. If the present trends continue it can be predicted that virtually the entire Riverine Plain—irrigated areas and dryland areas—will become a zone of regional groundwater discharge within 100 years. Similar land use changes have taken place across other parts of Victoria. The effect is the widespread appearance of land and stream salinization, both on the plains and in the highlands, in the latter instance the saline cycle of erosion is seen as a major factor in the modification of the Australian landscape.

The present day high incidence of salinity is not unique, but represents the most recent of a number of similar cycles of salinization which occurred naturally in the past and left their imprint on the land, the soils, the geomorphology and the groundwater chemistry. The physiographic evidence of earlier, naturally occurring cycles of higher water tables and salinization provides an important clue to the likely nature, direction and form of future groundwater discharge, with its own attendant salinity problems.

Throughout the study area there is marked contrast in the response of the regional aquifers to changes in surface water budget. During wetter periods, groundwater storages rapidly fill, and potentiometric heads and water tables may quickly rise. On the other hand, increased aridity, while causing almost instantaneous surface drying, has little immediate impact on the groundwater systems—the corresponding decline in water tables at these times is either slow or even nonexistent. Therefore, while high water tables and saline landscapes may readily arise under favourable hydrologic conditions, once this condition is attained it requires a very long and intense period of surface drying to reverse the process. For instance, the presence of prolonged and widespread lunette forming conditions following the regional drying out of the landscape in the late Pleistocene shows that the response of the groundwater systems to climatic change may lag thousands of years behind that of the surface systems.

While the response of the landscape to Quaternary climatic change is partly governed by the intensity of the climatic fluctuation, it is also strongly influenced by the pre-existing hydrologic character of the environment. In this respect, during times of increased aridity, the groundwater systems are a significant buffer to long-term environmental shifts within the drainage basin. The presence of certain groundwater influenced landforms, such as clay dunes, is not so much an indication of palaeoclimate as it is of palaeohydrology. These landforms often reflect hydrologic instability as the basin hydrology either re-adjusts to, or absorbs, the climatic fluctuation.

The Mallee

The locations of the groundwater discharge features in the Mallee are largely determined by structural and topographic influences. Of special significance is the presence of north trending faults, such as the Danyo and Tyrrell Faults, across the westerly flow paths of the major aquifers.

The discharge of groundwater to the surface creates its own distinct suite of landforms, best seen in the aeolian Mallee landscape, where discharge commonly occurs in the large discharge complexes (the boinkas), or in the many small salt lakes scattered throughout the Mallee. While most of the water in the salt lakes comes from precipitation, the bulk of the salt is derived from the groundwater systems, thereby imposing a large degree of chemical uniformity on the salt lakes of

northern Victoria. In the semiarid continental settings of northwestern Victoria, groundwater brines form from concentrated lake water which seeps into the underlying aquifer. The loss of salt from lakes into the aquifers commonly occurs under high lake regimes, when favourable hydrostatic head differences permit reflux of lake brines. Further salt loss occurs as the lake finally dries out, at which time the concentrated bittern enriched lake waters fall below the salt crust or the lake floor to become the uppermost part of the groundwater system. Once the lake is dry, the salt crust, if present is periodically permeated by a fluctuating water table, and those ions not already at levels of saturation in the shallow groundwater are redissolved from the salt crust and enter the groundwater regime. This hydrodynamic process explains the absence of significant amounts of magnesium and potassium minerals in the thin halite deposits of Victorian salt lakes. It also explains the relative scarcity of halite producing lakes despite the fact that their salinities commonly reach the halite saturation stage during drying.

In the Tyrrell Basin, hydrostatically induced reflux processes have produced a groundwater brine body with salinities ranging from 90,000 mg/l up to 290,000 mg/l. It covers an area of about 200 km², and occurs to a depth of about 70 m. The brines have formed within the last 32,000 years. The presence of dense groundwater infilling the aquifer beneath a lake has a major effect on the nature of lake–groundwater interactions, leading to the virtual capture of the regional groundwater flow system in the vicinity of the lake by means of up-flow along a Ghyben–Herzberg interface. On account of its greater density, the brine body may underflow an otherwise closing groundwater divide, and thereby open the lake basin to outflow via the groundwater systems. Similar high density groundwater brines are found beneath many of the larger salt lakes and boinkas of northwestern Victoria.

While groundwater discharge occurs at a number of points in the landscape, these discharge areas are not the final depositional point for salt carried by the regional groundwater flow system, but, instead, provide temporary storage sites—both at the surface as halite deposits and in the aquifers as brine bodies. The relatively short-lived conditions which favour salt accumulation within individual lake basins are eventually replaced by conditions conducive to outseepage into the aquifer, after which the salt continues downbasin towards the final continental repository—the Murray River. As a consequence, there are no significant thick salt deposits in northern Victoria, other than as groundwater brines.

In the Mallee, the regional groundwaters show marked uniformity and closely resemble seawater in both salinity and in ionic composition. The strong influence of regional groundwater flow on the salt lakes of northern Victoria is seen in their hydrochemistry, which invariably resembles a concentrated regional groundwater. The resulting uniform chemical character of the salt lakes enables the adoption of a chemical framework—based on the chloride and bromide ion concentrations in the lake waters. The chemical analysis of lake waters within this framework provides a guide to the nature of the annual cycle of halite precipitation and re-solution (if it occurs), and the paths taken during evaporitic concentration of lake waters; it also provides an insight into the interactions between salt lakes and the groundwater system.

In the lower regions of the drainage basin, groundwaters are commonly acid, and this has a significant influence on hydrochemical and biochemical processes in the discharge zones. Acidity in the unconfined Parilla Sand aquifer is caused by redox processes acting on iron and sulphur species, partly in response to climatically induced vertical water table movements during the Quaternary and Recent periods. The most significant change in water budgets observed in the late Quaternary sequences in the study area was the drying up of Lake Tyrrell at about 32,000 years BP, when lake water levels and regional water tables fell about 14 m.

In the Tyrrell Basin vertical redox zonation is paralleled by pH zonation, whereby acid groundwaters having a positive E_h are found closer to the water table, while neutral pH water having a lower E_h occurs at depth. At Lake Tyrrell acid springwater, with pH values ranging from about 4.0 to 2.6, outflows along the western margin of the lake and on the Tyrrell Creek salt flats to the south. The Tyrrell Creek for much of the time is a base-flow stream with a pH of about 4.

The acid springwater is derived from the regional groundwater. In contrast, neutral pH springwater occurs along the northeastern shoreline, derived from the more saline Timboram–Wahpool reflux brine generated in Lake Timboram. The differences in pH of the two inflowing waters are reflected in the hydrochemical processes active in the spring zone.

The high chlorinity and low pH of the regional groundwaters enable the ready transport of metal ions from upbasin source areas into the discharge zones. Where groundwater discharges around the perimeter of salt lakes, acid conditions in the spring zone are replaced by neutral conditions at the lake edge, thus providing an important zone for the precipitation of previously soluble ions. Iron oxide deposits of groundwater origin are a common feature of the shorelines of the larger discharge lakes, and this process may be responsible for many similar iron deposits and other metal ore deposits found elsewhere in continental settings. Indeed, the regional discharge zones in general and the spring zones of salt lakes in particular must be regarded as significant sites in continental settings for the concentration of metal ions carried by regional groundwaters. The groundwater discharge salt lake setting provides a simple model to explain certain of the syngenetic/epigenetic ore bodies where there is an association of red beds, evaporites and metal sulphides. Here, the source of metal ions may be either localized or disseminated throughout the aquifer system.

Finally, it is noted that the character of the groundwater systems in northern Victoria and their ongoing effects on the landscape are strongly influenced by a number of interacting factors which are themselves varyingly active through time. Of these factors, time itself is also significant in the flat Australian landscape where low hydraulic gradients produce long groundwater residence times. The climatic factor is of special importance since it determines total water availability. It subsequently controls the nature of the groundwater - surface water interactions which in turn strongly influences hydrochemistry and geomorphology. The nature of the interactions between the groundwater systems and surface waters, both determines in part, and is partially determined by, the positions and forms of the hydrological thresholds existing within a drainage basin. The positions of the thresholds fluctuate both temporally and areally in response to climatically induced variations in surface water availability.

Therefore, for a full understanding of the character of the groundwater systems, their evolution and their influence on surface processes, the groundwater and surface water systems must be regarded as aspects of a single integrated non steady-state flow system.

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This publication details the interaction between groundwater and surface systems in northern Victoria, Australia, as reflected in the hydrochemistry and hydrodynamics of the waters and the palaeohydrology and geomorphology of the landscape. The nature and form of regional groundwater discharge zones in semiarid landscapes is examined. The work details the high degree of hydrologic instability in the regional aquifers, and how increased aquifer recharge as a consequence of post European landuse changes has led to widespread salinization. A framework for salinity management at a regional level is included.

There is a detailed account of the hydrochemistry of salt lakes and the hydrodynamics of salt lake - aquifer interactions, including the development of groundwater brine bodies. The distribution and development of acid groundwaters is discussed together with their potential to mobilize and transport base metals to form syngenetic and groundwater epigenetic sulphide deposits in the vicinity of salt lakes.