Physical and chemical hydrogeology of the Otway Basin, southeast Australia

Angela L. Bush

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Declaration

This is to certify that

- (i) the thesis comprises only my original work towards the PhD;
- (ii) this work has benefited from a data exchange with my colleagues at SKM, who also completed a hydrogeological mapping project in Victoria in 2009;
- (iii) due acknowledgement has been made in the text to all other material used, and;
- (iv) the thesis is less than 100,000 words in length, exclusive of tables, maps, bibliographies and appendices.

Signed,

Angela Louise Bush

Abstract

The Otway Basin of southeast Australia is the subject of this thesis, which incorporates pre-existing geological, hydraulic and major element hydrogeological data with new isotope hydrogeochemical investigations. The region is an Upper Cretaceous–Tertiary basin, filled with siliciclastic and calcareous aquifers and aquitards and characterised by late volcanic activity, pervasive faulting and karstification.

As part of this study, an hydrogeological database is compiled for the Otway Basin region from existing distinct datasets from the states of Victoria and South Australia. Utilising this new resource, the data are reinterpreted into a 3D model of the hydrostratigraphy for the basin in *GoCAD*, and interpolated surfaces of hydraulic head and electrical conductivity are created for 5 aquifers/aquitards. The Victorian hydraulic head data is analysed for long term declining or inclining trends and hydrograph trend maps are created for different aquifer systems. The data are also compiled into representative cross sections of flow and chemical composition, with one section located in each of the three major sub-basins. The records of groundwater chemistry from the Victorian section of the Otway Basin are used to plot the relative concentration of major cations and anions for the main aquifers. More than 120 groundwater samples were taken for analysis of major and minor ion concentration and/or oxygen, hydrogen, carbon, strontium and chlorine isotope composition. These data are used to characterise the hydrogeochemical evolution of the groundwater and to identify the processes that the groundwater drives or experiences in the system.

The potentiometric maps and cross sections reveal the interconnected nature of the flow in all aquifers and the relationship between local and regional flow systems. Regional flow paths originate inland near basement highs or the basin margins. In the shallower aquifers they terminate at the coastline where the groundwater mixes with ocean water at a diffuse interface and density differences induce groundwater discharge at the land surface or the ocean floor. In the deeper confined aquifers, discharge is submarine via several possible mechanisms, which include: diffuse intergranular leakage to overlying units; flow along faults or volcanic conduits; and/or seepage directly to the ocean from exposed sections of the aquifer, e.g. in submarine canyons. These mechanisms may be operating up to 50 km offshore but the interface is currently migrating landward, which will result in a shortening of that estimated distance.

Local-scale flow lines are complex and may be oriented against the direction of regional coastward flow. Local hydraulic divides are often associated with volcanic eruption centres, which have elevated topography and relatively high hydraulic head, making them important recharge zones. These zones contain low salinity groundwater because infiltration is relatively rapid. Conversely, basalt flows that have developed clay horizons through weathering reduce drainage and allow significant evapotranspiration which concentrates the cyclic salts in solution. Many local flow systems discharge mainly via evapotranspiration, which acts again to concentrate the cyclic salts in solution. Other local discharge zones are rivers, creeks and lakes or lagoons that receive baseflow and seeps and springs associated with geological contacts or boundaries and faults. Evaporitic concentration of solutes in surface water bodies and shallow groundwater affects the quality of water recharging the underlying aquifers and aquitards. This quality has changed over the last 50,000 years or so due to fluctuations in climate and hence variation of the precipitation/evaporation ratio.

Stresses on the aquifers are climate fluctuations, sea level change, land use change and groundwater extraction. These stresses have resulted in the system being out of hydraulic equilibrium in many cases. Lags in response to these changes in boundary conditions are identified and/or hypothesised. In particular, the confined aquifer's response to sea level change could be subject to a lag in the order of millennia. The stress on an aquifer is often transferred to its adjacent units, in some cases inducing cross-formational leakage, which is possibly supported by radiocarbon dating evidence. The area of the Otway Ranges appears to have escaped the effects of stress to date because of its stable microclimate, its distance from the ocean and from groundwater extraction. Increase in demand on groundwater resources, development of geothermal, sequestration and hydrocarbon industries and future climate change may yet have a detrimental effect on the groundwater of the Otway Basin.

Isotopic composition of the groundwater confirms its meteoric origin and chlorine isotopes from several samples of the deep groundwater indicate that accumulation of solutes along the flow path is not due to diffusion or dissolution of connate salt. Thus, the salinity of the water is sourced from cyclic salts and solutes from water-rock interaction, both of which may be concentrated by evapotranspiration.

Water-rock interaction is dominated by dissolution of carbonates and weathering of silicates as a result of the surficial geology being dominated by calcarenite or limestone and young basalt. The volcanic activity has produced gas that has interacted with the groundwater, and continues to do so, fractionating oxygen, hydrogen and carbon isotopes and contributing fluorine, boron and sulphur to solution. The addition of volcanic CO_2 creates an uncommon situation for water-rock interaction, where continued dissolution of carbonate and silicate minerals along the deeper flow paths is demonstrated by the silicon/chloride ratios and strontium isotopic composition of the groundwater. These water-rock interaction processes, with the addition of cation exchange, are responsible for the development of a relatively fresh Na⁺HCO₃⁻ type water that is characteristic in parts of the deep aquifer.

The study confirms the existing hydrogeological understanding of the Otway Basin and forms new conclusions regarding the history of the groundwater and the processes of flow and chemical evolution by integrating numerous lines of evidence. Significant contributions of this work which improve current scientific knowledge include these findings: the maps and cross sections of hydraulic head and electrical conductivity reveal the connected nature of flow systems within all the aquifers and aquitards; geological features can induce discharge, e.g. at contacts or faults, and recharge, e.g. volcanic eruption centres; the changes to the surface drainage system as a result of the eruption of basalt flows have affected the water chemistry and flow systems in all the underlying units; there is a lag in aquifers' responses to sea level change and therefore future migration of the interface is expected regardless of further boundary changes; climate change has influenced surface water quality by changing the regional water balance, and therefore has affected groundwater quality; the discharge from the confined aquifer is submarine via various pathways, interaction between the groundwater and volcanic gas has occurred in the past and is ongoing, and consequently mineral dissolution persists at deep levels; the origin of high salinity of brackish groundwater in all Tertiary aquifers and aquitards is concentrated solutes from water rock interaction and cyclic deposition.

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Dedication

This work is dedicated to the traditional owners of the land of the Otway Basin: the Ngarrindjeri, Buandig, Gunditjmara, Girai wurrung, Djargurd wurrung, Gadubanud and the Gulidjan.

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Abbreviations

Abbrevaition	Meaning
BP	before present: years prior to the year 1950
DIC	dissolved inorganic carbon
EC	electrical conductivity
GMWL	global meteoric water line
HCO	Holocene climatic optimum
ITCZ	intertropical convergence zone
ka	kiloannum: thousands of years before present
LGM	last glacial maximum
LK	lower Cretaceous
LT	lower Tertiary
LMWL	local meteoric water line
Ma	megaannum: millions of years before present
MIS	marine isotope stage
ML	megalitres
PDB	Pee Dee belemnite
P/E ratio	precipitation–evaporation ratio
SGD	submarine groundwater discharge
SMOW	standard mean ocean water
SOI	southern oscillation index
SST	sea surface temperature
STA	sub-tropical anticyclonic belt
TDS	total dissolved solids
UK	upper Cretaceous
UT	upper Tertiary
WTF	watertable fluctuation

Chapter 1

Introduction

The Otway Basin is a diverse geological province which hosts valuable groundwater, hydrocarbon and geothermal resources both onshore and offshore. The volcanic rocks at the surface constitute an important building material and provide nutrient-rich soil that is beneficial for agriculture. The region is currently undergoing a dynamic transformation due to its increasing population and new and expanding ventures occurring in the area, such as pine and eucalypt plantations, hydrocarbon and geothermal exploration and CO_2 geosequestration. Such developments pose risks to, or may impose changes upon, the high quality groundwater resources of the basin. Also at risk are numerous ecosystems in which surface water-groundwater interaction plays a vital role, including: coastal lagoons, wetlands, salt and freshwater lakes, karst features and streams and rivers. Threats to these bodies of water, and to the quality and sustainability of the groundwater, also include land use change, natural and anthropogenic climate change and sea level change.

However, before these risks or changes can be quantified or managed, a greater understanding of the hydrogeological system is needed. This need is underscored by the dependence of many communities in the Otway Basin region upon groundwater resources. Groundwater has been extracted for irrigation and municipal supply since at least the 1950s. It has also been used for several decades by the natural gas and petroleum, dairy and smelting industries. Hence, there is a high demand for groundwater currently placed on the Otway Basin aquifers, which will increase as the population grows. Therefore, this study aims to provide the robust scientific framework which is currently lacking.

Firstly, this thesis addresses some hydrogeological phenomena of the Otway Basin which have been previously documented, but are not fully understood. For example, cones of depression in relation to groundwater extraction have been documented in the basin, but the effects of this development are not well constrained (SKM, 2005a). A major anticipated risk related to this is seawater intrusion, as much of the extraction occurs near the coast. Additionally, falling groundwater levels have been recorded in surficial aquifers in many parts of the study area and are connected to the drying of lakes (Ingram *et al.*, 1989; Adler & Lawrence, 2008). Groundwater extraction and climate change have been identified as possible causes of this change, but their effects on the deeper aquifers are largely unknown.

Second, the hydrogeology issues identified in the better-studied areas of Australia's southeast are considered as possibly also affecting the Otway Basin. The most relevant areas are the neighbouring sedimentary basins, the Gippsland Basin to the east and the Murray Basin to the north. In the coeval Gippsland Basin, the effect of extensive hydrocarbon extraction is currently evident in the strongly declining groundwater pressure in the aquifers (Hatton et al., 2004). This has created serious problems for irrigators and groundwater managers of the Gippsland Basin. As the Otway Basin is at an earlier stage of development than the Gippsland Basin, this study comes at a time when it is still possible to document the state of the basin's aquifers prior to extensive hydrocarbon extraction. In contrast, the major groundwater issue in the Murray Basin is the change in recharge rate associated with land use change. Increased recharge as a result of deforestation and irrigation has mobilised salt from the previously unsaturated zone and solutes have since been concentrated by evapotranspiration, causing devastating land and water salinisation (Allison et al., 1990; Jones et al., 1994; Evans, 1998; Herczeg et al., 2001; Wilson 2002). A better understanding of similar processes in the Otway Basin is needed, in addition to consideration of the possible effects of increasing land use for plantations.

Third, this study integrates information across the basin as a whole, whereas many previous studies have analysed only particular aquifers or discrete sub-basins. The first and only other major project that encompassed both the South Australian and Victorian sections of the Otway Basin was the Joint Bulletin published by these states' Geological Surveys (Wopfner and Douglas, 1971), which is a comprehensive account of the geological knowledge of the basin at the time. However, the hydrogeology chapter contained therein (p363-375) only addresses the Victorian section of the basin. Extensive work has been done to correlate the data sets of the two states through reports on the Border Zone required by the Groundwater Borders Agreement (1985). However, these reports pertain to a limited slice of the Otway Basin. The distinct lack of a "comprehensive description of the regional hydrogeology" was remarked upon by Leonard (2003); and this project aims to fill that void.

1.1 Objectives of the Study

The aim of this study is to understand and define the regional hydrogeology of the coastal Otway Basin by building a robust regional conceptual and 3D model which incorporates an understanding of physical and chemical processes. This is accomplished through the following specific objectives:

- to identify the hydrostratigraphic units and catalogue their lithological and hydraulic characteristics;
- to construct a digital 3D model of the hydrostratigraphic framework;
- to map the contemporary groundwater flow systems and recharge and discharge zones of the main aquifers, including submarine discharge zones;
- to understand the structural and lithofacies controls on groundwater flow and composition, encompassing the effects of Recent volcanism;
- to incorporate palaeohydrologic knowledge into this understanding and assess the effect of past and future sea level and climate change;
- to identify significant response in aquifer behaviour to land use change and groundwater abstraction; and
- to document the patterns and controls of the hydrochemical evolution of the groundwater that occurs via input of cyclic salt and water rock interaction.

1.2 Regional Setting of the Otway Basin

The east-west trending coastal Otway Basin has an area of approximately 150,000 km² onshore and offshore and occupies the southwestern coast of Victoria and the southeastern corner of South Australia (Laing *et al.*, 1989; Figure 1.1). The basin was formed by the continental rifting of Australia from Antarctica which began about 150 million years ago in the late Jurassic (Wopfner & Douglas, 1971). During rifting and subsidence, four major transgression and regression cycles occurred, resulting in the deposition of up to more than 4,000 m of marine and terrestrial sediments throughout the Cretaceous and the Tertiary (Bock & Glenie, 1965; Holdgate & Gallagher, 2003).

The sediments of the Otway Basin have been variably classified into hydrostratigraphic units by different authors (see Chapter 2). In this thesis, the units (aquifers and aquitards) are identified and a configuration of the hydrostratigraphy is developed by drawing on current stratigraphic and lithologic knowledge and by correlating units across the state border.

The sand, gravel and limestone aquifers are separated and confined by marl and mudstone aquitards and have lithofacies variations both in a seaward-landward (north-south) and in an east-west orientation. In addition, a complex, outcropping volcanic aquifer



Figure 1.1. Location of the Otway Basin study area in southwestern Victoria and southeastern South Australia. Adapted from Harrington *et al.*, (1999).

system covers much of the Tertiary sediment. The Otway Basin aquifers and aquitards have been deformed by the extension during rifting and by inversion events throughout its evolution. Uplift during inversion events has created basement highs that demark embayments, or sub-basins, within the basin.

The study region has a Mediterranean climate with rain occurring mainly in the winter, and evapotranspiration dominating in the hot summer. The region is used widely for agriculture and forestry, with pasture and pine plantations having replaced most (90%) of the native vegetation of the area (Love *et al.*, 1993).

The geomorphology of the Otway Basin region is wonderfully diverse, encompassing low-lying basalt plains, alternating subparallel dune-and-swale systems, a mountain range more than 550 m above sealevel, meandering river valleys or creek gullies, hummocky fractured rock landscapes of the volcanic "stony rises", steep sea cliffs, sandy beaches of low relief, undulating karst features including sinkholes, and volcanic craters and cones. Figure 1.2 shows a map of the topographic relief of the above onshore features (horizontal resolution of approx. 90 m; http://seamless.usgs.gov/), in addition to the bathymetry offshore (resolution of 1 km; Geoscience Australia, 2006). The bathymetry clearly shows the extent of the continental shelf and the submarine canyons that incise the continental slope.

The groundwater resources in the area are extensively developed for agricultural irrigation and municipal supply (Leonard *et al.*, 1983; Blake, 1989; Love *et al.*, 1993). Most agricultural and pastoral properties have private bores for extracting stock or irrigation water from the shallow aquifers. Municipal supplies, especially for the larger towns, typically come from the deeper confined aquifers.

The extent of the Otway Basin falls over the boundaries of two Victorian catchment management authorities: the Glenelg-Hopkins CMA and the Corangamite CMA. Across the state border in South Australia, the aquifers are confined within the South East Catchment Water Management Board Region (Figure 1.3). These authorities influence policy in relation to catchment health, which is affected by groundwater-surface water interaction. In addition to catchment management, there are also water resource management bodies in the Otway Basin region. For example, the Victorian section of the basin is administered by Southern Rural Water and several water supply protection areas (WSPA) exist within the basin (http://gmu.geomatic.com.au/Default.aspx). Management bodies are responsible for allocating licences for groundwater extraction in a way which ensures the sustainability of the groundwater resource, as required under the Water Act of 1989 (http://www.austlii.edu.au/au/legis/vic/consol_act/wa198983/). Along the state border, the groundwater resources are jointly managed between the states under the Groundwater Border Agreement Act, 1985 (http://www.austlii.edu.au/au/legis/sa/consol_act/gaa1985291/).

1.3 Thesis Outline

Each chapter of this thesis focuses on a particular aspect of the basin's geology, hydrology or hydrogeology and contains an appraisal of existing research, an introduction to the approach and analytical methods employed, a discussion of the results obtained and a synthesis of the significance of the findings. The concluding chapter then presents the integrated findings of all the chapters in a concise summary.

Chapter 2 describes the geology and geological history of the study area. The geology is explored in terms of the basin structure, sedimentation and lithologies. In particular, the hydrostratigraphy is introduced and defined in terms of its lithologic and structural characteristics. The depositional and structural histories of the area are integrated, providing a comprehensive physical framework for the study.

Chapter 3 is a summary of the modern climate of the Otway Basin region, including







Figure 1.3. The boundaries of the Catchment Management Regions in relation to the study area.

an explanation of how typical weather patterns are maintained or changed. It includes an extensive literature review concentrating on the palaeoclimate of the region. This review integrates geological, meteorological and hydrological evidence to reconstruct a history of the climate and hydrological balance of the basin which informs the hydrogeological investigations in subsequent chapters. Glacio-eustatic cylces are an important climate control and are therefore discussed within the palaeoclimate history. In addition, eustatic fluctuations are relevant in that they act to change the boundary conditions of coastal aquifers.

Chapter 4 presents the physical hydrogeology of the basin. The contemporary flow systems of the hydrostratigraphic layers are depicted in maps and described. The regional patterns of recharge and discharge are constrained and are related to processes of groundwater-surface water interaction. Offshore discharge is discussed, and by connection, groundwater-marine water interaction through the use of oceanographic data, bathymetry data and eustatic records. The controls on groundwater flow of aquifer lithofacies, structure and geomorphology are characterised. Assuming steady state conditions, Darcy's Law is applied for the calculation of indicative groundwater velocity in two of the hydrostratigraphic units.

The flow of the groundwater in the aquifers of the Otway Basin is further constrained

by use of representative cross sections in Chapter 5. The transient state of the aquifers is illuminated by analysis of hydrographs and this work is used to show that several stresses are currently affecting the groundwater. The influence of factors such as long term climate change, sea level fluctuations, land use change and abstraction are discussed.

In Chapter 6, the major dissolved constituents of the groundwater of the Otway Basin are analysed. Two main types of parameters are used in this analysis: the first type is electrical conductivity (EC) measurements; the second is major and minor anion and cation concentrations. Maps of the groundwater EC for each of the hydrostratigraphic units are presented. Original hydrogeochemistry results from this study are integrated with the historical major ion records for the Victorian section of the basin. Contemporary along-flow groundwater chemistry trends are linked to water-rock interaction and surface processes, such as cyclic salt input, CO_2 sources and evapotranspiration, and are compared to previously published findings.

The isotope hydrogeochemistry of the groundwater is then presented in Chapter 7. Specific analytes discussed in the text include stable, radiogenic and radioactive isotopes, namely: ¹³C, ²H, ¹⁸O, ⁸⁶Sr, ⁸⁷Sr, radiocarbon and ³⁶Cl. The isotope geochemistry is used to characterise the recharge signature of the groundwater in different regions, determine residence times for groundwater using piston flow assumptions and trace processes of water-rock interaction. A further application of these analytes is a deeper understanding of the control of the past climate on the groundwater quality of the confined aquifers.

Finally, the concluding chapter of this thesis (Chapter 8) presents a brief synthesis of the findings from each chapter and examines their significance in terms of the study objectives (Section 1.1). This chapter thus outlines the regional framework of the Otway Basin provided by this study. This framework will be of benefit to those seeking to assess both the future risks to the groundwater resources of the Otway Basin and the most sustainable path for their management.

All the data used in this study are compiled in digital appendices attached to this thesis. This is to facilitate fast and easy comparisons between this study and others. The data are also provided digitally to make them accessible for use by groundwater resource stakeholders in the region, be they managers of water resources, managers of ecology or land managers. The physical 3D model constructed as part of this study is also appended to the thesis in digital form.

Chapter 2

Geology, Hydrostratigraphy and Basin History

2.1 Introduction

This chapter presents an interpretation of publicly available geological data alongside a synthesis of findings of other authors which relate to the lithology, environment of deposition and structural history of the basin. In this way, an overall understanding of the physical framework within which the hydrogeology of the basin must be interpreted is achieved. Primary investigation of the geology beyond field observations of permeability, weathering characteristics and porosity was not attempted in this study as the main focus of this research is the occurrence, behaviour and character of the groundwater of the basin.

A detailed understanding of the geology of the region is important to a groundwater study such as this for two reasons: a) to supply a physical framework of the distribution of the hydrostratigraphic layers and the hydraulic properties associated with lithofacies, and b) to provide histories of the depositional environment and geomorphology of the region. All these factors have influences on the groundwater flow system, including, for example, shaping the chemistry of the connate water, limiting the flux of subsequent recharge and developing discharge zones. This chapter is organised into the following sequential topics: methodology, hydrostratigraphy, basin structure and tectonic history.

The hydrostratigraphic units are listed in the legend of Figure 2.1 and are described in chronological order within the text. The onshore lateral extent of the Tertiary sediments can be seen in Figure 2.2, which also shows the onshore surficial extent of the Pliocene-Quaternary Volcanic Aquifer. The Lower Cretaceous Basement extends below the majority of the distribution of the Tertiary sediment and is the primary hydraulic basement. As mentioned in the introduction, the tectonic history of the basin resulted in buckling and uplift of the basement and other layers, producing somewhat discrete embayments within the Otway Basin. These zones, too, are depicted in Figure 2.2 and provide a context within which to discuss the lateral heterogeneity within the basin.

The current distribution of outcrop of the hydrostratigraphic units is shown in the geolgical map of Figure 2.1. This map was developed by adjusting geological maps from the Geological Survey of Victoria (2003) and Geological Survey of South Australia (2004) to the framework devised in this study. For continuity the colour scheme used to represent the units in this map is employed in subsequent figures. In contrast to most geological maps, where colour relates primarily to rock type and age, the colours in this map relate to the hydraulic characteristics of the rocks. Blue and green relate to aquifers whereas aquitards are represented by red or orange. The Pliocene-Quaternary Volcanic Aquifer is represented in pink in accordance with convention for igneous rocks; its dominance in the regions of the Tyrendarra and Port Campbell Embayments notable from Figure 2.1. In contrast, the rarity of outcrop of older aquifers is also visible in this map.

2.2 Methods

2.2.1 Hydrostratigraphy

The hydrostratigraphic framework used in this study was constructed by correlation of lithofacies of similar hydraulic and geochemical properties that are coniguous or share a geological contact, be it lateral or vertical. Before the 3D framework could be constructed, a comprehensive literature investigation, collating both published and unpublished reports, was necessary. The framework construction involved some grouping of stratigraphic formations together with their neighbours. At the same time, other layers were separated from their parent formation in order to be grouped with lithologically similar formations. The grouping and separation was achieved by using established lithofacies descriptions from the various sources (referenced within the text).

The research into each unit was concentrated on the characteristics important to groundwater, i.e. the descriptions of grainsize, mineral type (e.g. clay content and geochemical composition), porosity (including secondary karst porosity), cementation and consolidation were considered crucial to delegation of aquifer or aquitard status. While descriptions of palaentology, age and environment of deposition are provided, they are important for this study mostly in the way that they influence the characteristics above. For example, many of the aquifer boundaries are diachronous.

In addition, treatment of particular formations or lithofacies by previous authors in hydrogeological literature as aquifers or aquitards also influenced the construction of the hydrostratigraphic framework for this study. Examples of conflicting opinions in regards








to a geological unit's ability to act as an effective aquifer or aquitard are provided where appropriate, but are usually the result of lateral lithofacies changes within a unit studied in disparate regions, rather than a distinct lack of agreement between authors.

Aspects of geomorphology were also researched at the same time as the geological history and the lithology. The geomorphology is discussed in the text where relevant but the implications for groundwater are mostly reserved for subsequent chapters.

2.2.2 GIS and Physical Models

GIS data from the Geoscience Victoria (The Department of Primary Industries) contained information on surface geology and provided the lithological logs for the Victorian section of the Otway Basin used to create the 3D hydrostratigraphic model. This information had coordinates in eastings and northings (zone 54) and in latitude and longitude (geographic data) in the AGD66 datum (geodetic). In this case, the grid coordinates were converted to geographic coordinates in the more modern datum, GDA94. The geological logs obtained from Department of Primary Industries and Resources South Australia (PIRSA) had location information in both grid and geographic coordinates, in the GDA94 datum. There was no need to adjust these spatial coordinates. Well completion reports (WCRs) for gas and petroleum exploration and production were also incorporated into the geologic database. The location of all the 966 wells used in the model are plotted in Figure 2.3.

The individual logs from the two states were processed in such a way that they were converted into one cohesive database. The logs were adjusted such that, rather than detailing the depths of the stratigraphic layers at a particular location, they presented the elevations (in mAHD) of the tops and bottoms of the hydrostratigraphic units (defined and detailed below). Through this process, the lithofacies changes and nomenclature differences within the stratigraphic units and across the expanse of the basin were taken into account and presented in a form that is appropriate for a groundwater study.

Having been sourced from individual reports (from drillers logs to full-scale WCRs), the quality control of the source data is relatively diverse. Therefore, it was necessary to analyse the data for anomalies and incongruencies. Some logs clearly did not conform with neighbouring ones and were hence discounted. The reasons for these spurious logs are probably varied, but are believed to include: a) incorrect location information; b) incorrect depth intervals (due to human error during data entry); c) poor stratigraphic analysis/correlation (perhaps due to lack of time, lost returns or insufficient geological analysis); and d) incorrect or imprecise elevation information, especially for older records. The case of poor elevation data was overcome in some instances by extracting ground elevations for the given borehole coordinates from an SRTM digital elevation model (DEM) with a spatial resolution of 95 m (the dataset is provided free online by USGS). This



Figure 2.3. The location of all geological logs used in this study and the various cross sections, both from this study and sourced from published works (Moore *et al.*, 2000; Geoscience Australia, 2005). Cross sections are presented later in the chapter. Bathymetry is represented by 1000 m contours.

method was also useful for identifying spurious data, made obvious in cases where there was a large error between the provided elevation and the extracted elevation.

Once the database was finalised it was imported to the 3D program (GOCAD 2.1.6), and the hydrostratigraphic model was constructed by interpolating between the logs for various horizons. The layers mapped were specifically the thickness of each hydrostratigraphic unit in every log and the elevations of the tops of the units in each log. The method of interpolation was inverse distance weighted (IDW). The main function of interpolation in a regional study is to "fill in the gaps" in areas of sparse data. In some cases, this has resulted in a better understanding of the regional geology. In others, it does not represent certain facets of what we already know about the geology, especially where faulting has a strong influence on the sediment layers. The advantage of using GIS and other methods of 3D modelling is predominantly in the clarity of three dimensional visualisation; which can otherwise be problematic.

2.3 Hydrostratigraphy

Geological investigation of the stratigraphy of the Otway Basin has, by Australian standards, a long history. Woods (1860; 1862) was one of the first to study the South Australian region and wrote of the volcanism at Mt Gambier, the stranded beach dune systems and the widespread Tertiary limestones. Similarly, in Victoria, early documentation was dominated by the geology of the extinct volcanoes and the seaside cliffs of the eastern Otway Basin (e.g. Bonwick, 1858; Murray, 1877). In addition, the search for coal and gold was motivation for early geological investigation (e.g. Murray, 1880). Correlation and comparison of the Tertiary strata from various basins or regions was developed or attempted by several authors (e.g. Woods, 1865; Tate & Dennant, 1895). This task became easier with more drilling and interpretation of boreholes into the 20^{th} century, partly driven by speculation on the basin's potential for petroleum and groundwater resources (Wade, 1915).

Since these early works, several key publications have been thorough in defining and categorising the stratigraphic units of the Otway Basin (e.g. Sprigg, 1952; Boutakoff, 1963; Bowler, 1963; Bock & Glenie, 1965; Wopfner & Douglas, 1971). Bock and Glenie (1965) constrained the limits of four distinct eustatic cycles within which are further shorter transgressions and regressions. The main stratigraphic formations of the basin, their environment of deposition and their distribution within the hydrostratigraphic system are represented in Table 2.1. Although each hydrostratigraphic unit is described below, at times, Table 2.1, complied from a number of sources, may be the more detailed catalogue within this thesis of the lithological characteristics of the units. The basin sediments include sandstone, mudstone, limestone, marl, gravel and sand, deposited in a diverse range of environments (Bock & Glenie, 1965; Love *et al.*, 1993; Monteil *et al.*, 2004).

There remain differences in stratigraphic nomenclature in Victoria and South Australia (e.g. Parker, 1993), however, when correlated, the units create a cohesive and contiguous hydrogeological system. In Figure 2.4 the seaward-landward (roughly equivalent to SSW-NNE) distribution patterns of the aquifers and aquitards are represented to a general degree (mostly representative of the eastern part of the basin). Other seaward-landward depictions of the hydrogeological framework are in Figures 2.5, 2.6 and 2.7, which are

Table 2.1. Hydrostratigraphy and lithology of the Otway Basin. T and R refer to transgressive and regressive marine phases respectively. Geological time periods and unit thicknesses are not to scale. Note that the chronological relationships between the units, especially the Quaternary units, is more complicated than what is represented here. Adapted from Bock & Glenie (1965), Tickell *et al.* (1992), Holdgate & Gallagher (2003) and Geoscience Australia (2007).

Era	Period	Epoch	Group	Phase	Hydrostratigraphic Unit	Formation Name	Depositional Environment	Lithology
Cainozoic		Pleistocene			Plio-Pleistocene Heterogeneous Aquifer	Lowan Sand & Simpson Sand	Aeolian dune and interdunal alluvial	Unconsolidated, ferrugenous, medium to fine quartz sand, some calcareous and gypsiferous fragments, weakly modified by pedogenesis.
						Malanganee Sand	Aeolian	Fine to medium grained quartz sand derived from other Pleistocene units
						Padthaway Formation	Lacustrine, lagoonal, colluvial	Unconsolidated sand, silt, sandy clay, peat, marl & freshwater limestone.
	aternary					Bridgewater Formation	Coastal barrier dunes, shallow sub-tidal, aeolian	Bioclastic, cross-bedded calcarenite, rich in silica and some heavy minerals, interbedded with palaeosols and capped by calcarenite.
	Qui					Coomandook Formation	Intertidal, coastal	Partly lithified, grey, cream, fossiliferous, medium-grained quartz sand & calcareous sand; characteristically contains lithic clasts and skeletal debris derived from older limestone.
					Pliocene-Quaternary	Tuff rings	Terrestrial, volcanic	ash, scoria, bombs, lapilli
						Scoria cones		ash, scoria, bombs, lapilli and calcareous fragments
					r olouino r iquilor	Lava flows		olivine basalt
						Whalers Bluff Formation	Coastal, shallow marine	Sandy limestone, calcarenite, shell beds, marl.
		Pliocene			Plio-Pleistocene Heterogeneous Aquifer	Parilla Sand & Dorodong Sand	Aeolian, lacustine, fluvial and	Micaceous sand, quartz sand and sandy clay with some rare fossils.
						Hanson Plain Sand	Marine-to-non-marine transition	Quartz Sand and clayey sand
						Managha al Maduat Canad	zone	Very course-grained quartz sand, calcareous sand, conglomerate,
						Moorabool Vladuct Sand	Marine and non-marine	sand & silt. Weakly computed calcarenite (limestone), fine grain, guartz rich
		Miocene	g	т	Upper Tertiary	Port Campbell Limestone	Deep marine	minor clayey silt
			hunc	т	Carbonate Aquifer	Gambier Limestone (west)	Shallow marine shelf	Bioclastic, karstified, at times dolomitised limestone, rich in bryozoa and with basal glauconitic marl.
			test	(R)	Upper Tertiary	Gellibrand Marl (east)	Deep marine (shallowing)	Calcareous silty clay and clayey silt, minor shelly calcarenite beds,
		e	Hey	т	Upper Tertiary	Clifton Formation	Shallow marine	Glauconitic calcarenite (limestone), minor quartz and limonite sand,
		cene		-	Carbonate Aquifer			some beds well cemented
		Oligo	a Gp	т	Upper Tertiary Aquitard	Narrawaturk Marl Demon's Bluff Formation (east)	Deep Marine	Calcarenous mudstone, minor calcarenite, local calcareous and glauconite pellets.
		e	Nirranda	т			Shallow marine	Silt, fine sand, clay, clayey sand, carbonaceous, pyretic.
	iary					Mepunga Formation	Deep marine (shallowing)	Quartz sand, minor limonite and glauconite, localised cementation (calcite), otherwise unconsolidated. Alternating layers of carbonaceous clayey silt, including basal shale (Sturgess Point Mb)
	Tert					Demon's Bluff Formation (east)		
		ocen			Lower Tertiary Sandy Aquifer	Older Volcanics	Sills and dykes	Olivine bearing gabbro and basalt, deeply weathered intrusions are
		Palaeocene				Dilwyn Formation Eastern View Formation (east)	uay-non.	
				R			Deep marine to shallow marine	Sandy clay, silt, carbonaceous, burrowed, quartz sand, minor gravel and shale, massive to moderately bedded
			_				Olivine bearing gabbro and basalt, deeply weathered intrusions are	
			Wangerrip Gp	_		Older Volcanics	Sills and dykes	clay-rich.
				Т	Lower Tertiary	Pember Mudstone	Deep marine to shallow marine	Siny day, dayey sin, minor line quartz sand, carbonaceous, limonitic, burrowed mudstone
				т	Aquitaro	Pebble Point Formation	Deep marine to shallow marine	Dolomitic, quartz-rich sandstone, minor glauconite pellets, quartz pebbles, well bedded, interbedded with carbonaceous clay and basal mudstone
				R		Moomowroong Sand Member	Marine-to-non-marine transition	Quartz sand, minor clay, fine grained mica, friable, minor gravel
				R	Upper Cretaceous Aquifer	Wiridjil Formation	Non marine	Quartz gravel, sand, minor pebble layers, minor clay, carbonaceous, friable, minor volcanic and metamorphic pebbles
Mesozoic	1			R	(upper part)	Timboon Sand	Marine Transition zone	Quartz sand and gravel, carbonaceous mudstone, generally unconsolidated
			Sherbrook Gp	R		Paaratte Formation	Deep and shallow marine	Quartz sandstone, siltstone, mudstone, minor pebble beds,
		Late		т		Nullawarre Greensand	Shallow marine	Quartz sandstone, glauconite pellets, weakly cemented, fine-to-
				т	Upper Cretaceous	Belfast Mudstone	Deep marine (shallowing)	coarse-grained
					Aquitard	Deliast Mudstolle	Deep manne (snanowing)	Siltstone mudstone sandstone thin coal beds quartz grains
	snoeous			т	Upper Cretaceous Aquifer	Flaxman Formation	Deep marine (shallowing)	ooliths, glauconite pellets, dolomite, siderite, minor phosphate, basal mudstone
	ret			т	(lower part)	Waare Formation	to non marine	mudstone, minor coal, fine to coarse
		Early	Otway Gp		Lower Cretaceous Basement	Eumeralla Formation	Volcanolithic sandstone, siltstone, mudstone, mud clast conglomerate, feldspar, quartz, coal, consolidated, well bedded, medium to fine Sandstone with minor, fossiliferous siltstone and coal. Shale, siltstone with minor fine-grained sandstone. Quartz and feldspar sandstone, minor siltstone, mudstone and coa medium to coarse, kaolinite, garnet, mica Volcaniciastic sand and olivine basalt interbedded with minor sandstone, siltstone and shale	Volcanolithic sandstone, siltstone, mudstone, mud clast conglomerate, feldspar, quartz, coal, consolidated, well bedded, medium to fine
	1				(not addressed in this study)	Katnook Sandstone		
	1					Pretty Hill Sandstone		Quartz and feldspar sandstone, minor siltstone, mudstone and coal,
	1							
	-	\square				Casterton Formation		sandstone, siltstone and shale
alaeozoic			Bedrock		Bedrock	Metamorphosed sediments and plutonic rocks		

cross sections from the basin margin inland to the offshore basin edge in the south or southwest. The cross sections' locations are marked in Figure 2.3; each one represents one of the major embayments of the basin, the Gambier, Tyrendarra and Port Campbell Embayments. The cross sections give a general overview of the relationships between the hydrostratigraphic layers described in detail below. In addition, they provide an insight into the extent and form of the horizons mapped in the physical 3D model because they were produced from the same dataset.

By contrasting the cross sections from the different Embayments, the lateral (NW-SE trending) changes in thickness of the hydrostratigraphic layers are evident. These variations add another dimension of complexity to the basin framework. As investigated below, they are mainly due to variable depositional environments throughout the basin's history and they act as controls on groundwater flow in some regions. As an overview of the lateral relationships, Figure 2.8 is presented. It is a fourth cross section, its location also shown in Figure 2.3, and is oriented subparallel to the modern Otway Basin coastline. The morphological differences between the Embayments is more striking in this cross section than it is from comparing the previous sections. In addition to the Warrnambool High at bore number 99488, the basement highs that mark the northwest and southeast margins of the basin, namely, the Padthaway Ridge and the Otway Ranges (refer to Figure 2.2), are identifiable in the shallowing of the depth to basement at either end of the section.



Figure 2.4. Stratigraphic formations and hydrostratigraphic units of the Otway Basin, their ages, relative average thicknesses and the relative sea level change during deposition. The uppermost formations, the Pliocene-Quaternary Volcanic Aquifer and the Plio-Pleistocene Heterogeneous Aquifer are not depicted here. Adapted from Bock & Glenie (1965), Leslie (1966), Love *et al.* (1993) and Krassay *et al.* (2004).















Figure 2.8. A coastline-oriented cross section of the Otway Basin hydrostratigraphic units.

2.3.1 Lower Cretaceous Basement

The Lower Cretaceous Basement is represented by the Eumeralla Formation of the Otway Group, which extends underneath the entire Basin and is the dominant rock formation of the Otway Ranges (Figure 2.1). The morphology of this pervasive basement feature is illustrated in the map of Figure 2.9, which shows the structure contours of the top of the Eumeralla Formation. The map also shows the regions where the Eumeralla Formation and other Palaeozoic Basement rocks crop out. Together, these components depict the lateral and depth hydraulic extents of the Otway Basin. Further, the basin structure is visible in this map, revealing, not just the greater thickness of sediment towards offshore, but also the basement highs that were depicted in Figure 2.2, namely, the Dartmoor Ridge, the Warrnambool High and the Barongarook High, which partially separates the Barwon Downs Embayment from the rest of the basin.

The Eumeralla Formation equivalent was first recognised in the Gambier Embayment in a bore drilled at Robe, SA and was classified as Jurassic, as were all the Otway Group occurrences in Victoria at the time. The Robe occurrence was identified by correlation with a sequence near Casterton, VIC (Sprigg, 1952). The beds comprise argillaceous and arkosic sandstone with minor bituminous coal and are stained green due to chlorite cementation. They are characterised by preservation of feldspar and poor sorting, indicating rapid deposition and burial in a freshwater environment, probably estuarine (Sprigg, 1952).

2.3.2 Upper Cretaceous Aquitard

This aquitard is mostly composed of the Belfast Mudstone. This unit effectively constitutes the basement of the hydrogeologic system in some areas of the basin as it can occasionally directly overly the Eumeralla Formation which is the ultimate hydraulic basement formation (Section 2.3.1). This only occurs where the Upper Cretaceous Aquifer is absent between the two aquitards, see Figure 2.8. Therefore, basement is defined as the top of the Upper Cretaceous Aquitard, or where the lower part of the Upper Cretaceous Aquifer is present, the Lower Cretaceous Basement. The Belfast Mudstone was deposited in a deep marine environment with the environment shallowing towards the top of the sequence (Bock & Glenie, 1965). It comprises carbonaceous, glauconitic and pyritic mudstone with minor quartz sand (Tickell *et al.* 1992).

2.3.3 Upper Cretaceous Aquifer

This Aquifer group is usually confined above by the Lower Tertiary Aquitard, although where this is missing, a direct contact with the Lower Tertiary Sand Aquifer can occur. It





is usually underlain by the Upper Cretaceous Aquitard, which could at times be considered hydraulic basement. However, it can be interbedded with this unit, creating complexity and necessitating a further definition of the deeper Basement (Section 2.3.1). The main reason for the inter-fingering of the Upper Cretaceous Aquifer and the Upper Cretaceous Aquitard is that the Upper Cretaceous Aquifer comprises several lithostratigraphic units and is subdivided into the upper and lower parts. The upper part includes some lower Tertiary units, despite the name of this aquifer system. These are the Pebble Point Formation, the Moomowroong Formation and the Wiridjil Formation. The upper part of the Upper Cretaceous Aquifer includes from the Cretaceous: the Timboon Sand, the Paaratte Formation and the Nullawarre Greensand. The lower part includes the Flaxman Formation and the Waarre Formation.

Lower Part

Waarre Formation

The Waarre Formation is currently developed and explored offshore as a host rock for oil and gas sourced from the underlying Eumeralla Formation (Lower Cretaceous Basement). It is mostly sandstone intercalated with occasional carbonaceous mudstone, deposited in an open marine and shallow marine setting of the mid-Cretaceous transgression (Bock & Glenie, 1965; Tickell *et al.*, 1992).

Flaxman Formation

The Flaxman Formation comprises deep marine sediment, dominated by ferruginous, dolomitic sandstone and siltstone, although the basal unit is mudstone (Bock & Glenie, 1965). Minor mudstone and thin coal beds are also present in this unit. Clasts include quartz grains, ooliths, glauconite pellets, and minor phosphate (Tickell *et al.*, 1992).

Upper Part

Nullawarre Greensand

The Nullawarre Greensand contains poorly sorted quartz sand and glauconite pellets, with minor berthierine (Boyd, 2005). It was deposted in a shallow marine environment (Tickell *et al.*, 1992).

Paaratte Formation

The Paaratte Formation is a deep to shallowing marine complex which consists of poorly-sorted, cemented sediment ranging from pebbles to shale (Bock & Glenie, 1965). It is carbonaceous and contains minor glauconite.

Timboon Sand

The Timboon Sand is mostly unconsolidated gravel and quartz sand intercalated with

carbonaceous mudstone (Tickell *et al.*, 1992). It represents the littoral and paralic zone during a regression, with some coal and minor plant remains indicating terrestrial proximity.

Moomowroong Formation & Wiridjil Formation

These formations were previously recognised as the Wiridjil Gravel Member and the Moomowroong Sand Member of the Pebble Point Formation (see below). They are thin lenses of friable sediment deposited in a variety of littoral to non-marine environments. Lithology and mineralogy is diverse, (including quartz, carbonate, metamorphic and volcanic clasts), but grainsize is generally large (pebble, gravel and sand) with the exception of some fine-grained mica and clay.

The above formations were mapped as a single hydrostratigraphic unit and the resultant isopach map can be seen in Figure 2.10. Although the Pebble Point Formation (described below) is partially included in the Upper Cretaceous Aquifer, it is not included in the aquifer thickness shown in Figure 2.10. Its relatively thin, lensoidal nature, partly a product of its deposition after a major unconformity, was the reason for its exclusion from the map. However, from the dashed line marking the combined extent of the Upper Cretaceous Aquifer, it can be seen that a significant part of the Upper Cretaceous Aquifer in the Port Campbell Embayment and the Barwon Downs Embayment is composed only of the Pebble Point Formation. The overall distribution of the upper part of the Upper Cretaceous Aquifer is characterised by a seawards deepening and thickening which is typical of most of the basin's units.

Pebble Point Formation

The Pebble Point Formation comprises ferruginous sandstone and some conglomerates. Clasts are mostly quartz, often chlorite or glauconite pellets or ooliths, sometimes dolomitic and rarely fossiliferous (Holdgate & Gallagher, 2003). Fossils, including forams, pollens and macrofossils (including ostracods) provide an early- to middle- Palaeocene age for this formation and an transgressive shallow marine palaeoenvironment with influences from the sublittoral and estuarine environments (Holdgate & Gallagher, 2003). Three well defined fining-upwards sequences exist in the Pebble Point Formation within the Port Campbell Embayment. Each sequence grades from a conglomerate basal unit through very coarse sand to fine sand and carbonaceous silt and clay (Bock & Glenie, 1965). Some of the finer-grained lithofacies are excluded from the Lower Tertiary Aquifer (see Section 2.3.4).

The Pebble Point Formation is equivalent to the once-named Bahgallah Fomation, the basal unit of the Knight Group in the Gambier Embayment. In general, the Bahgallah Formation is an unconsolidated, dark green and brown chamositic, oolitic medium- to coarse-grained sand with occasional silt or carbonaceous beds deposited in an open marine





environment (Kenley, 1971; Ludbrook, 1971).

In the Gambier Embayment, the Pebble Point Formation (or the Bahgallah Formation) is divided into three different lithofacies, although it is not certain whether the these correlate laterally with the fining-upwards sequences in the Port Campbell Embayment (Holdgate & Gallagher, 2003). The basal lithofacies is a pelletal and oolitic mudstone (Holdgate & Gallagher, 2003). The middle lithofacies is thin and comprises ferruginous yellow to brown sand and interbedded white silt and quartz gravel (Kenley, 1971). The fossil assemblage is sparse and includes shell fragments, burrows, wood (near the base), sharks teeth and coral. The upper facies of the Bahgallah Formation comprises sandy clay and sand, yellow- to red-brown in colour, containing chamositic ooliths and pellets and interbedded with rare white silt beds (Kenley, 1971). These strata are more fossiliferous than those underlying and present many species which are abundant in the Pebble Point Formation (e.g. *Cucullaea psephea, Lahillia australica* and foraminifera) proving the correlation between these units (Kenley, 1971).

2.3.4 Lower Tertiary Aquitard

The main stratigraphic component of the Lower Tertiary Aquitard is the Pember Mudstone. The Pember Mudstone was initially classified as a member of the Dilwyn Formation (Glenie, 1971) and is still sometimes referred to as the Pember Mudstone Member (Holdgate & Gallagher, 2003). However, it is now considered a formation of the Wangerrip Group (Smith *et al.*, 2003; McGowran *et al.*, 2004). The Pember Mudstone usually conformably overlies the Pebble Point Formation and comprises grey and pale brown mudstone, shale and siltstone with some interbedded sandstone (Holdgate & Gallagher, 2003), containing burrows and deposited in a deep marine to shallow marine environment (Tickell *et al.*, 1992). Mineralogy is micaceous, pyritic, carbonaceous and occasionally glauconitic (Holdgate & Gallagher, 2003). The Lower Tertiary Aquitard also contains certain lithofacies of the Pebble Point Formation that have low hydraulic conductivity, e.g. clay and mudstone (Tickell *et al.*, 1992); a full description of the Pebble Point Formation is given above.

The extent and thickness of the Lower Tertiary Aquitard is mapped in Figure 2.11. It is interesting to note that the inland extent of the Lower Tertiary Aquitard is not as great as that of the under- and over-lying aquifers (Figures 2.10 & 2.12 respectively). The implication is that, in almost all the marginal zones of the basin, the Lower Tertiary Sandy and and the Upper Cretaceous Aquifers are direct contact and hydraulic connection may exist between the groundwater therein.





2.3.5 Lower Tertiary Sandy Aquifer

The Lower Tertiary Sandy Aquifer is mainly constituted by the Dilwyn Formation and its lateral equivalents. The Lower Tertiary Sandy Aquifer also comprises the Eastern View Formation and locally, the previously-named "Older Volcanics". The Lower Tertiary Sandy Aquifer is also locally in hydraulic connection with parts of the Mepunga Formation that have adequate hydraulic conductivity. However, generally the Mepunga Formation is classed as an Aquitard. For a more detailed description, see Section 2.3.6. The Lower Tertiary Sandy Aquifer is generally confined above by marl of the Upper Tertiary Aquitard (Upper Tertiary Aquitard) and below by mudstone and clay of the Lower Tertiary Aquitard (Lower Tertiary Aquitard). However, direct contact with either the Upper Cretaceous Aquifer or the Upper Tertiary Carbonate Aquifer exists in some parts of the basin.

The greatest accumulation of the Lower Tertiary Sandy Aquifer is up to c. 900 m thick and is situated in the Tyrendarra Embayment, close to Portland and Heywood (Figure 2.12). In contrast, the aquifer is relatively thin in the Port Campbell Embayment, the Barwon Downs Embayment and the majority of the Gambier Embayment. Across the basin, the maximum thickness is reached approximately at the location of the current coastline and offshore from this zone the Lower Tertiary Sandy Aquifer thins and presumably lenses out. The current known offshore extent, not limited to this isopach map, but better constrained by seismic surveys, indicates that the unit is contiguous to the break of the continental shelf in some places. Where this is the case, it is usually truncated by the shelf edge or incised by the deep submarine canyons whose heads reside there (see Section 2.4).

Also mapped for the Lower Tertiary Sandy Aquifer are the structure contours of its upper surface (Figure 2.13). The morphology of the aquifer mimics that of the Basement, illustrated in Figure 2.9. The Port Campbell Embayment is visible as a deep trough and the Gambier Embayment is relatively shallow by comparison with a dome structure north of Mt Gambier, near Mt Burr. The manifestation of the Ferguson Hill anticline and the Barwon Downs graben to its southeast are apparent in the structure contours in the Barwon Downs Embayment. The structural features in this area and the bottlenecklike narrowing of the extent of the Lower Tertiary Sandy Aquifer act to separate the groundwater within the embayments on either side.

Dilwyn Formation & Equivalents

The Dilwyn Formation is part of the Wangerrip Group, bounded above by a mid-Eocene unconformity, and comprises carbonaceous and quartz-rich sands, interbedded with clay, silt and minor gravel; and was deposited in a deep to shallow marine setting during an









Eocene regression (Bock & Glenie, 1965; Blake, 1980). The Dilwyn Formation of the central Otway Basin was deposited in a mostly paralic environment and comprises some argillaceous beds and also red and yellow ferruginous sand (Glenie, 1971).

The Dilwyn Formation is composed of sandstone interbedded with more minor siltstone and claystone in transgression-regression coarsening-upwards cycles. Its age ranges from middle Palaeocene to Early Eocene and it contains rare fossils (usually within the fine-grained strata) preserved by arenaceous and carbonaceous minerals (Holdgate and Gallagher, 2003). The Dilwyn Formation conformably overlies the Pember Mudstone in most regions. Within the Port Campbell Embayment the lithology is dominated by sand which grades into its eastern equivalent, the Eastern View Formation, in the Torquay Embayment. In these regions, the units exhibit coal seams which become increasingly thick to the east. The thinning of the Dilwyn Formation over the regional basement highs, such as the Warrnambool High, was caused by uplift and erosion prior to the deposition of the overlying Nirranda Group (Holdgate and Gallagher, 2003). Offshore seismic interpretation revealed a largely southwards-progradational geometry of the lower Dilwyn Formation. The source of this prolific clastic marine sedimentation is possibly terrigenous and may have been due to a previous course of the River Murray whose mouth was much further south than today (Holdgate and Gallager, 2003).

The western equivalent of the Wangerrip Group is the Knight Group. The name of the Knight Group is now obsolete, as are most of the names of the formations and members contained therein. However, the original names are used here for clarity and are placed in the context of being equivalents of the Dilwyn Formation, therefore conveying their place in the overall stratigraphic framework. The Knight Group was deposited in a shallow, warm paralic environment and comprises sand, gravel, clay, minor brown coal, brown mudstone in upper parts with clay and Fe-rich minerals (Sprigg, 1952). Fossils in this Group include: Cyclammina & Victoriella (forams), sharks' teeth, bryozoa (rare) and plant matter (Sprigg, 1952). Within this Group, the Dartmoor Formation and Tartwaup Formation (overlying the Bahgallah Formation) are the equivalents to the Dilwyn Formation. The Dartmoor Formation overlies the Bahgallah Formation (Middle Palaeocene) with a conformable, gradational boundary. It has an unconformable relationship with the overlying Nelson Formation (Kenley, 1971).

The Dartmoor Formation is typically a grey to brown silty sand with fine quartz grains and occasional dolomitisation (Ludbrook, 1971). Micaceous dark grey silt and fine sand are present at the base of the Dartmoor Formation (Kenley, 1971). Interbedded with the dominant quartz sand are black clay lenses (similar to the basal unit) which contain fossils (e.g. gastropods) and are otherwise rare in the rest of the Formation (Kenley, 1971). The upper beds include quartz sand and ferruginous (limonite cement) sandstone with interbedded shale and minor coal, carbonaceous silt or clay, dolomitic sand, etc. Mica, pyrite and organic matter are common throughout. Grainsize is very variable, from medium to very coarse sand to occasional gravel (Kenley, 1971). Several species of the benthic foraminifera *Cyclammina* are well preserved but forams are not abundant (Ludbrook, 1971). The sequence was deposited in a mostly paralic environment, with occasional input from littoral marine and deltaic or estuarine environments (Kenley, 1971).

The Tartwaup Formation comprises brown coarse sand with abundant quartz, pyritic quartz and some silt. Near the top of the sequence shark teeth, mollusk shell fragments and *Cyclammina* are found (Ludbrook, 1971). Interspersed within the Tartwaup Formation is the Burrungule Member which is composed of carbonaceous, micaceous clay and silt and has a distincive foraminifera assemblage. Overlying the Middle Eocene Tartwaup Formation is the similarly-aged Kongorong Sand which is a characteristic sequence of coarse quartz sand with limonite staining (Ludbrook, 1971). Although it is not part of the Knight Group, the Kongorong sand is included as part of the Lower Tertiary Sandy Aquifer and probably correlate to the top of the Dilwyn Formation.

Eastern View Formation

The Eastern View formation is the equivalent of the Dilwyn Formation in the Barwon Downs Graben and the Torquay Basin (not addressed in this study). It is similar to the marginal marine deposits of the Wangerrip Group further west but is markedly different in the presence of substantial intercalated coal seams.

Volcanics

The "Older Volcanics" consist mostly of basalt of varying geochemistry (e.g. titanaugite basalt and olivine basalt) but also include coarse grained intrusions (Kenley, 1971). It is often intercalated with the Dartmoor Formation or within the Knight Group. These basalt, lamprophyre and dolerite deposits are also found interbedded within the Dilwyn Formation (Holdgate and Gallagher, 2003). From K-Ar dating it has been shown that the distribution of these rocks throughout the various embayments occurred as distinct events spread over many millions of years, from about 59.1 Ma to 17 Ma (Holdgate and Gallagher, 2003).

2.3.6 Upper Tertiary Aquitard

The Upper Tertiary Aquitard comprises the Gellibrand Marl, the Greenways Member, the Narrawaturk Marl, the Mepunga Formation, the Demons Bluff Formation and the Burrungule Member. It is the main confining unit for the underlying Lower Tertiary Sandy Aquifer (Lower Tertiary Sandy Aquifer). Figure 2.14 is an isopach map for the combined thickness of these units. It also includes the thickness of the Clifton Formation, which is not an aquitard, but is confined between the Narrawaturk Marl and the Gellibrand Marl and therefore included in the isopach map. The most notable feature of this map is that the aquitard is less than 50 m thick throughout much of the Gambier Embayment. This is in contrast to development of thicknesses in excess of 400 m in all the other Embayments. The difference is due to east-west lateral facies changes within the relevant stratigraphic formations which result in aquifers grading into aquitards or *vice versa*.

There may be a further component of the aquitard offshore, in the form of Miocene canyon fill, although the hydraulic characteristics of this lithofacies is unclear and therefore it may rightly belong in the Upper Tertiary Carbonate Aquifer (Upper Tertiary Carbonate Aquifer). A generation of Miocene sub-marine canyons were identified on the shelf of the Otway Basin using seismic line interpretation and were shown to incise the Upper Tertiary Carbonate Aquifer and the Upper Tertiary Aquitard (Leach & Wallace, 2001). Subsequent Miocene canyon fill therefore unconformably overlies the Upper Tertiary Aquitard in the region of the canyons. It is difficult to classify the canyon fill sediment into a hydrostratigraphic category as little is known about its lithology. Leach and Wallace (2001) proposed, based on seismic indicators, that the fill was likely to have a higher carbonate content than the juxtaposed Miocene slope facies, which is most likely marl of the Upper Tertiary Aquitard. Therefore, the canyon fill could possibly be part of the Upper Tertiary Aquitard or the Upper Tertiary Carbonate Aquifer.

Burrungule Member

The Burrungule Member is contained within the upper-most section of the Dilwyn Formation in the Region of the Gambier Embayment and the Dartmoor Ridge (Holdgate & Gallagher, 2003). The Burrungule Member is mostly carbonaceous fossiliferous silt and mud with minor sandstone. The fossil assemblage contains carbonaceous foraminifera and marine shells (Holdgate & Gallagher, 2003). The Sturgess Point Member of the Nirranda Group is of a similar lithology and is also incorporated into the Upper Tertiary Aquitard. The Burrungule Member and the Sturgess Point Member were deposited during the Early Eocene and the Middle Eocene respectively (Holdgate & Gallagher, 2003).

Mepunga Formation

The Mepunga Formation, while categorised as an aquitard by Love *et al.*, (1993) is considered by other authors to be an aquifer hydraulically connected to the Dilwyn Formation of the Lower Tertiary Sandy Aquifer (Bock & Glenie, 1965; Leonard *et al.*, 1983). This





is a case of spatial heterogeneity within the Mepunga Formation's lithology, being more marly in the Gambier and Tyrendarra Embayments and sandy in the Port Campbell Embayment. For the propose of this study, the Mepunga formation is taken as part of the Upper Tertiary Aquitard.

The Mepunga Formation unconformably overlies the Dilwyn Formation (Lower Tertiary Sandy Aquifer) in much of the basin. The Mepunga Formation comprises a mix of carbonate and clastic sediments. It thins towards the offshore margin and over the structural basement highs. The palaeoenvironment has been interpreted from rare fossils (such as planktonic foraminifera and miliolinids) and varies from inner/mid-shelf in the east to mid/outer-shelf in the west (Holdgate & Gallagher, 2003). Variation of palaeoenvironment is reflected in changes in lithofacies from east to west. Variations in lithofacies are also perceptible from north to south, where southern sections are carbonaceous open marine and in the north clastic shoreline facies dominate (Holdgate & Gallagher, 2003).

Demons Bluff Formation

Time equivalent of the main Anglesea coal seam and the Narrawaturk Marl, the Demons Bluff Formation was identified in the Torquay Basin, and is only present in the eastern section of the Otway Basin (Witebsky *et al.*, 1995). It comprises a range of mostly clastic sediment, including fine, clayey sand, silt and clay, with minor carbonaceous sediment (Tickell *et al.*, 1992). The Demons Bluff Formation presents a more restricted marine influence than that of the Narrawaturk Formation (Abele *et al.*, 1988) and is considered to represent a shallow marine to tidal flat environment (Blake, 1980).

Narrawaturk Marl

The Narrawaturk Formation is dominated by marine carbonates of varying depths (Witebsky *et al.*, 1995). As with the underlying Mepunga Formation, north-south-oriented facies changes are present. Within the basal strata of the Formation, glauconitic and shelly mudstone and siltstone of the north represent low energy marginal marine environments. They grade to pyritic marl in the south, which represents an outer marine shelf environment (Holdgate & Gallagher, 2003).

Palaeoenvironments of the upper part mimic those of the lower part, becoming deeper water environments in the offshore direction (south). The upper parts of the Narrawaturk Marl comprise richly fossiliferous dark grey marl (sometimes greenish due to the presence of glauconite), silty marl, calcareous mudstone and muddy limestone (Abele *et al.*, 1988). These strata are weakly bedded, widely bioturbated and intercalated with minor ferrugenous sandstone or calcarenite (Witebsky *et al.*, 1995). East-west facies differences also exist within the Narrawaturk Formation. In the west it is dominated by marly limestone, in the east, marl and mudstone (Holdgate & Gallagher, 2003).

In the region of the state border, in the east of the Gambier Embayment, the Narrawaturk Formation is often dolomitised. While the Narrawaturk Formation has been recognised in some wells in the Gambier Embayment, the formation generally lenses out and the equivalent rocks are attributed to the lower portions of the Gambier Limestone (Holdgate & Gallagher, 2003).

Greenways Member

The Greenways Member is the Early Oligocene basal marl of the Gambier Limestone, which is a component of the western portion of the Upper Tertiary Carbonate Aquifer. The Upper Tertiary Aquitard underlies the Upper Tertiary Carbonate Aquifer. The Gellibrand Marl underlies the Upper Tertiary Carbonate Aquifer in the east, however it lenses out towards the west. There, the Greenways Member (coeval with portions of Gellibrand Marl) and the Narrawaturk Marl underly the Upper Tertiary Carbonate Aquifer and confine the Lower Tertiary Sandy Aquifer.

The Gambier Limestone was first subdivided by Lindsay (1967) and subsequently the Greenways Member was formally described by White (1995). The Greenways Member is a grey-green marl, rich in bryozoa with abundant chert and some limestone layers (Li *et al.*, 2000). The characteristic colour is due to the presence of glauconite, especially in the basal strata. The Greenways Member is sometimes absent from the stratigraphic sequence and is replaced by the Compton Conglomerate; this is thought to be due to tectonic uplift and associated erosion (Ludbrook, 1971).

The uppermost strata of the previously-named Lacepede Formation correlate to the Greenways Member. The Lacepede Formation contains laminated carbonaceous clay beds and is largely equivalent to the Narrawaturk Marl (Kenley, 1971; Abele *et al.*, 1988). The lithology of the upper parts of the Lacepede Formation is a fossiliferous, shelly marl, stained green from glauconite (Ludbrook, 1971). However, the Lacepede Formation has variable lithology and can also contain quartz and muscovite (Ludbrook, 1971).

In yet another variation of nomenclature, now superceded, the Greenways Member of the Gambier Limestone may be equivalent to the basal Nelson Formation of the Glenelg Group (Kenley, 1971). The Nelson Formation is described by Kenley (1971) as comprising unconsolidated and cemented sandstone of variable grainsize with occasional conglomerate. Ooliths and pellets are common and some clasts are sourced from the Dartmoor Formation (Dilwyn Formation, Lower Tertiary Sandy Aquifer), including quartz grains. Cementing minerals include iron-bearing minerals and carbonates, for example, dolomite, siderite, ankerite, calcite and pyrite. This unit often grades into nodular limestone and then into the Gambier Limestone. The description appears to fit that of the Greenways Member, as it, too, has a gradational contact with the overlying limestone and the Lacepede Formation shares the trait of containing quartz clasts.

Gellibrand Marl

The Gellibrand Marl is a thick, mostly homogeneous unit which directly underlies the Upper Tertiary Carbonate Aquifer (see below) in most of the central and eastern Basin. Generally, it grades into more limestone-dominated lithofacies in the west and is eventually replaced by the Gambier Limestone, which has a basal marl member. This lensing-out is signified by the diochronous contact between the Gellibrand Marl and the overlying Port Campbell Limestone of the Upper Tertiary Carbonate Aquifer. According to biostratigraphy, the contact occurs in progressively older sediment towards the west (Abele *et al.*, 1988), signifying that the upper portions of the Marl are absent there. The Gellibrand Marl also thins or becomes absent over some basement highs due to structural highs at the time of deposition (Witebsky *et al.*, 1995).

The Gellibrand Marl comprises dominant marl, with rare interbedding of calcarenite and chalk (Holdgate & Gallagher, 2003). It also contains some calcareous beds of siltstone and sandstone (Witebsky *et al.*, 1995). Bock & Glenie (1965) described the Gellibrand Marl, which was previously subdivided, as a single unit. However, there remain slight differences between the upper and lower sections in the east of the basin. The base of the Gellibrand Marl in the east is a thin (c. 6 m thick) fossiliferous clay. Above the basal sequence, the lower part of the Formation comprises blue-grey fossiliferous clay (Abele *et al.*, 1988). The upper part comprises a similar blue clay, but more calcareous and sometimes yellow, with charateristic burrowing and glauconite (Abele *et al.*, 1988). Within the upper strata are thin beds of limestone which bear concretions. In the west, the Formation is dominated by clay and marl, although a darker grey and more glauconitic. Bryozoa is abundant some thin layers of limestone are interbedded with the clay and marl (Abele *et al.*, 1988).

The marl was deposited during a Late Oligocene to Early Miocene transgression (Witebsky *et al.*, 1995). The fossil assemblage is dominated by pelecypods, bryozoa, and gastropods; foraminifera are also present (Abele *et al.*, 1988). These abundant and diverse foraminiferal assemblages denote a low-energy marine palaeoenvironment of variable water depth (Holdgate & Gallagher, 2003).

2.3.7 Upper Tertiary Carbonate Aquifer

The stratigraphic formations included in the Upper Tertiary Carbonate Aquifer are the Port Campbell Limestone and the Clifton Formation in the east, and the Gambier Limestone in the west. The stratigraphic relationship between these units is complex, largely due to lateral (east-west) facies changes. For example, the two limestones are not coeval, but there may be a gradational contact close to the Dartmoor Ridge between the youngest sections of the Gambier Limestone to the west and the oldest sections of the Port Campbell Limestone of the east (McGowran et al., 2004; this study, Figure 2.8). Around the Dartmoor Ridge, the limestone is very thin (e.g. between bores 307616 and 69962, Figure 2.8) and little hydraulic connection is expected across the boundary. The Gellibrand Marl, which is pervasive in Victoria, is further evidence of this complexity. It is attributed to the Upper Tertiary Aquitard in the east, however, in the far west of the basin, the Gellibrand Marl is absent because, near the State border, it grades to a limestone lithology belonging to the Gambier Limestone (Holdgate & Gallagher, 2003). Similarly, what is described as the Clifton Formation in the East is a second sub-unit of the Gambier Limestone in the west. The Clifton Formation has variable lithology and distribution, both outlined below, and is almost everywhere in Victoria separated from the limestone of the Upper Tertiary Carbonate Aquifer by the Upper Tertiary Aquitard.

Therefore, the Gambier Sub-basin's Upper Tertiary Carbonate consists mainly of the Gambier Limestone, whereas the other sub-basins have two components of the aquifer, the Clifton Formation and the Port Campbell Limestone, which are separated by an aquitard. The thickness of the limestone components of the aquifer has been mapped across the region (Figure 2.15). Two zones encompassed by a 0 m contour are visible in the map, one in the region of Dartmoor and the other over the dome structure north of Mt Gambier previously recognised in Figure 2.13. These zones may not be completely void of the Upper Tertiary Carbonate Aquifer but they certainly represent areas where it has thinned to significantly less than 30 m thick and where its extent may be patchy or lensoidal. The thinning of the limestone is due to erosion associated with tectonic uplift and/or regression (Sprigg, 1952; Ludbrook, 1971).

It is also shown in Figure 2.15 that the Upper Tertiary Carbonate Aquifer is absent from much of the Port Campbell Embayment and the entire Barwon Downs Embayment. As a consequence, the surficial aquifers (see Sections 2.3.8 & 2.3.9) rest unconformably and directly on the Upper Tertiary Aquitard in those regions.





Clifton Formation

The Clifton Formation type section is situated at the sea cliffs near Princetown, in the Port Campbell Embayment as first identified by G. Baker (1944) (Glenie, 1971; p211). Sometimes the Clifton Formation is hydraulically disconnected from the overlying limestone portion of the aquifer by the Gellibrand Marl (the Upper Tertiary Aquitard). It is debated whether or not the Clifton Formation is classed as an aquifer (refs, Bennetts etc). The lithology is dominantly glauconitic and sandy limestone (Tickell *et al.*, 1992). However, marine derived dolomite has been documented by Nicholaides (1997).

The Clifton Formation has a variable lithology comprising limestone, sandy limestone and sandy marl with occasional nodule horizons rich in phosphate and limonite (Holdgate & Gallagher, 2003). The environments in which these sediments were deposited were marine, ranging from middle shelf in the south, through to inner shelf and then to paralic in the north (Tickell *et al.*, 1992). Coarse grained clastic components of the paralic facies may be terrigenous, possibly sourced from the fluvial systems now represented as the "deep leads" (Holdgate & Gallagher, 2003; although other authors maintain that these fluvial systems contain Pliocene sediments and therefore were not active at the time of depositon of the Clifton Formation).

There are strata in South Australia time-equivalent to the Clifton Formation, namely the Camelback Member of the Gambier Limestone (described below). However, because of the independent development of the stratigraphy in the two states, these coeval units were not correlated until recently. Gallagher & Holdgate (2000) used biostratigraphic evidence from several wells to infer a correlation between the Clifton Formation and the Camelback Member despite lateral lithofacies changes. Hydrostratigraphically speaking, the most important difference between these two components of the Upper Tertiary Carbonate Aquifer is that the Camelback Member conformably underlies another part of the Upper Tertiary Carbonate Aquifer (the Green Point Member), whereas the Clifton Formation is overlain by the confining Upper Tertiary Aquitard. This relationship is illustrated in Figure 2.8.

Gambier Limestone

The Gambier Limestone was deposited in the Gambier Embayment in the Miocene, coincident with the development of its deeper water equivalent from the Upper Tertiary Aquitard, the Gellibrand Marl further to the east. It is composed of grey or cream coloured calcarenite with abundant bryozoa and other fossils and can be intercalated with thin marl and clay layers or can exhibit dolomitised zones or chert nodules (Kenley, 1971; Li *et al.*, 2000). The dolomite formed during the Miocene, shortly after deposition of the limestone; the diagenetic fluid was a mixture of oceanic water and meteroic water (James, *et al.*, 1993). The diagenesis led to greater cementation and hence lithification of the limestone in the south than the in north (Love *et al.*, 1993).

As previously mentioned, the subdivisions of the Gambier Limestone were formally described as discrete members by White (1996). The basal member, the Greenways Member, is part of the Upper Tertiary Aquitard and has therefore been described above. The middle and upper members are the Camelback and Green Point Members respectively. Although landward-seaward facies changes exist within the Gambier Limestone (Spring, 1952), the member subdivisions adequately describe the hydrostratigraphy.

The Camelback Member was formally described by White (1995) from its type section in the drillhole Greenways 1, located approximately 30 km east of Robe, in the Gambier Embayment. It is mostly contained within the Late Oligocene (White, 1996; Li *et al.*, 2000). Its lithology is dominated by bryozoan limestone with some marl (Li *et al.*, 2000) and pink or brown coloured dolomitised zones in the upper section (Li *et al.*, 2000; J. Lawson, pers. comm., 2007). The Camelback member is intensely karstified, which adds an important secondary porosity to the already significant primary porosity. For this reason, the unit has high transmissivity and is extensively developed for irrigation (J. Lawson, pers. comm., 2007). However, the high yield is not reliable everywhere, due to the fact that the karstification, although pervasive, is not regionally interconnected (Love *et al.*, 1993).

The Green Point Member was deposited in a variably cool to warm water environment in the Late Oligocene and Early Miocene (White, 1996; Li *et al.*, 2000). It is comprised of bryozoal limestone with a varying abundance of chert nodules representing more trophic environments and intercalated with minor marl sections (Ludbrook, 1971; Li *et al.*, 2000). The Green Point Member is also used for rural groundwater supply, but has significantly less porosity of karst origin than the Camelback Member. It is considered to be the time equivalent of

Port Campbell Limestone

The Port Campbell Limestone has a conformable, sometimes gradational, relationship with the underlying Gellibrand Marl (Abele *et al.*, 1988) of the Upper Tertiary Aquitard. On a broad scale, the Limestone was deposited in a neritic marine environment during a middle to late Miocene transgression (Bock & Glenie, 1965). However, on a finer scale, several glacio-eustatic cycles are discernible. Three regressions and two transgressions span the period of deposition and, therefore, the lithofacies represented within the Formation range from outer- to inner-shelf (Holdgate & Gallagher, 2003). The last phase of deposition of the limestone was followed by a regression during the Pliocene which allowed extensive subaerial erosion (Bock & Glenie, 1965). This formed a karst surface over much of the top of the Port Campbell Limestone which is now an unconformity where it is overlain by younger deposits. Contacts with the Bridgewater Formation and the Whalers Bluff Formation, both members of the Plio-Pleistocene Heterogeneous Aquifer, have been recorded (Abele *et al.*, 1988).

The Port Campbell Limestone consists of fossiliferous, often pure, sometimes marly limestone, chalky in many locations with a white to grey colour and includes rare chert (Bock & Glenie, 1965; Abele *et al.*, 1988). The limestone has variable cementation and can be clastic or bioclastic (Glenie *et al.*, 1968). Glauconite-bearing sequences within the Port Campbell Limestone associated with thin basal mudstone layers (Glenie *et al.*, 1968) may be representative of a similar environment as that in which the Greenways Member of the Upper Tertiary Aquitard was deposited. Although karstification is not as pervasively developed as in the Gambier Limestone, fractures and joints within the Port Campbell Limestone have been widened by carbonate dissolution and form a secondary porosity to compliment the primary intergranular porosity (Shugg, 1981).

2.3.8 Pliocene-Quaternary Volcanic Aquifer

Previously called the Newer Volcanics, those igneous extrusions of the Western District Province that erupted after 5 Ma constitute the Pliocene-Quaternary Volcanic Aquifer (Price et al., 2003). The Pliocene-Quaternary Volcanic Aquifer is often interbedded with Plio-Pleistocene Heterogeneous Aquifer sediments and the groundwater hosted therein may have some effective hydraulic connectivity, too. However, the Pliocene-Quaternary Volcanic Aquifer is conceptualised as an aquifer system distinct from other Quaternary deposits in this thesis in response to the unique geochemical and geomorphological features of the various igneous lithofacies. Previous authors have considered the volcanic deposits as constituting more than one aquifer system (e.g. Coram, 1996; Bennetts, 2005), however, this is beyond the scope of this study, which incorporates many deeper units on a more regional scale. The Pliocene-Quaternary Volcanic Aquifer has numerous lithofacies variations within it which control very strongly its variable hydraulic characteristics (Thompson, 1972). For example, the diverse hydraulic conductivities and porosities of ash, tuff, scoria, and basalt must all be taken into account for this aquifer. The geomorphology of the Pliocene-Quaternary Volcanic Aquifer is also an important factor that influences the surface water drainage regime (Joyce, 1988), and in turn, groundwater flow. The geomorphology is especially important considering the fact that the volcanic extrusions cover a large proportion of the basin's surface, influencing recharge immensely (Figure 2.1).

The recent Western District Province extrusions were traditionally divided into two

phases of eruption: the first phase from the early Pliocene (c. 4 Ma) and the second phase which began in the Pleistocene and continued to recent times (Thompson, 1972). The first phase has subsequently been divided into two groups, 4.6 Ma to 3 Ma which comprises predominantly tholeiitic basalts and 3 Ma to 1.8 Ma which contained more diverse rock types (Price *et al.*, 2003). Generally, the first phase consisted of thin basalt sheets (no more than 30 m thick) which flowed over the existing landscape and locally filled in fluvial channels, effectively burying the Pliocene channel sands, forming the so-called "deep leads". These olivine basalts are vesicular and have an anomalously high Ca content, presumably associated with their contact with and eruption though the underlying Tertiary carbonates (Thompson, 1972). Sprigg (1952) suggested that the first phase of the volcanics was mostly the result of fissure flow, with limited vent formation following this.

In much of the basin, Pleistocene sediments where deposited in the hiatus between the eruption phases which is part of the reason for the interlayering of the Plio-Pleistocene Heterogeneous Aquifer and the Pliocene-Quaternary Volcanic Aquifer mentioned earlier. For example, in the Gambier Embayment coastal dune deposits lie on top of early basalt flows (Sprigg, 1952).

The second, or Quaternary, phase of eruption comprised a more diverse range of styles, resulting in more basalt flows, scoria cones, maar tuffs and the "stony rise" deposits (Joyce, 1975). The pervasive presence of maars formed during this phase illustrates that the magmatism was more explosive than the previous phase. This may be due to magmatic gas escape or heating of groundwater by magma bodies resulting in explosive steam.

The isopach map of the Pliocene-Quaternary Volcanic Aquifer (Figure 2.16), shows the distribution of aquifer thickness in the Otway Basin, gleaned from the log data, overlain by a representation of the general extent of the Western District Province. The general extent of the volcanics was mapped by Moore and Wong (2002) and this was adjusted for Figure 2.16 with the aid of total magnetic intensity images (TMI) for Victoria (GSV, 2003) and South Australia (GSSA, 2004). As can be seen in this map, certain areas of known volcanic distribution have unknown thickness, such as the offshore areas. This is largely due to the lack of offshore wells with returns from the uppermost 50 m of section. In addition, certain areas, such as the northeast region, were beyond the spatial limits of the log data used and hence only the extent of the aquifer is mapped in that region.

As mentioned previously, the diverse lithofacies of the Pliocene-Quaternary Volcanic Aquifer, and the resultant heterogeneous geomorphology, have a strong influence on both the surface hydrology and groundwater recharge mechanisms. The physical features of the Pliocene-Quaternary Volcanic Aquifer are visible in Figure 2.16, where areas of thin distribution often translate to low-lying regions where large lakes (e.g. Lake Corangamite) have now formed, or where small peaks in thickness represent eruption centres that can act as preferential pathways for groundwater recharge. Lithofacies at eruption centres are usually tuff, ash or scoria, all with high porosity and efficient drainage characteristics. The majority of the "Western Plains", however, is covered by relatively flat basalt flows, represented in the isopach map by regions of widely spaced contours.

Price *et al.* (1997) presented whole rock major element data (XRF), minor element data and Sr isotope data for various basalt flows of the Western District Province. General ranges of some major elements of the basalts are the following weight percentages: SiO₂ 45–55; Al₂O₃ 13–16; FeO_{total} 8–13; MgO 5–12; CaO 7–10; Na₂O 2–5. Five studies of the petrology and geochemistry of the volcanics were collated into a representative table of results presented in Price *et al.* (2003). Values correspond quite closely to the above figures. Due to the large surface area exposed and the mineral content typical of a basalt, e.g. olivine, pyroxene and plagioclase, weathering is extensive and is often characterised by the development of a clay rich in montmorillonite (Joyce, 1975).

2.3.9 Plio-Pleistocene Heterogeneous Aquifer

The Plio-Pleistocene Heterogeneous Aquifer comprises many recent sediments and constitutes the relatively thin covering in much of the basin over the more substantial Upper Tertiary Carbonate Aquifer. The unit has been subdivided here by lithology into Subaquifer One, a collection of porous sands, calcarenites and gravels, and Sub-aquifer Two, consisting mostly of recent clay-rich deposits. Due to the complex and sometimes fragmented nature of the Plio-Pleistocene Heterogeneous Aquifer, and the lack of detail in the upper part of many of the logs, no isopach map was attempted. However, the spatial distribution of the two sub-aquifers is visible in Figure 2.1. It can be seen there, especially in the western region of the basin, that outcrops of Sub-aquifers One and Two are aligned in a coastal orientation, alternating between dune ridges (sandy units) and swales (clayey units). Also visible in Figure 2.1 are the alluvial and lacustrine deposits associated with the lakes of the Western Volcanic Plains in the eastern part of the basin.

Additionally, there is an offshore component of this aquifer system, the Pliocene canyon fill sediments of buried canyons found under the continental shelf. Leach and Wallace (2001) identified the deposits using seismic line and magnetic imagery interpretation and showed that, due to the incision of the canyons into the Upper Tertiary Carbonate Aquifer, the Pliocene canyon fill sediments (the Plio-Pleistocene Heterogeneous Aquifer) often unconformably overly the Upper Tertiary Aquitard. They proposed that the magnetic signature of the Pliocene canyon fill could imply a silicic or clastic mineralogy and sediment was possibly terrigenous, originating from the same sources as the Loxton and Parilla Sands.




Sub-aquifer One

Two pervasive dune-building sediment types, the siliceous Pliocene sands and the calcarenites of the Pleistocene, are incorporated in Sub-aquifer One of the Plio-Pleistocene Heterogeneous Aquifer. The Pliocene sands are composed mostly of medium to fine grained quartz sand and minor clayey sand, deposited in a shallow marine and marginal marine environment as the sea level regressed. Similarly, the Pleistocene deposits also document the change in position of the coastline due to a fluctuating sea level. However, rather than the sandy deposits preceding them, the Pleistocene dunes are mostly carbonate. In some areas, the Pliocene and Pleistocene components are separated by the extrusions of the Pliocene-Quaternary Volcanic Aquifer (Pliocene-Quaternary Volcanic Aquifer), which increases the complexity of the hydrostratigraphic model. Unnamed Recent sandy alluvial and colluvial sediments are also included in Sub-aquifer One. Some of the stratigraphic units are outlined below.

The Dorodong Sand unconformably overlies the Gambier Limestone of the Upper Tertiary Carbonate Aquifer (Kenley, 1971). It comprises a basal gravel below ferruginous fine grained sand and sandstone which is micaceous and very sparsely fossiliferous. Its distribution extends at least throughout the Casterton and Dartmoor regions and also north and west of Casterton. The Dundas laterite soil profiles are found within the Dorodong Sands. Low-relief strandline formations are present in this Miocene–Pliocene unit (Holdgate & Gallagher, 2003) and the quartz sand of the dune ridges are consolidated with a limonitecemented cap. It is proposed that the ridge deposits are a lateral equivalent of the Parilla Sand of the Murray Basin and that the remainder of the Dorodong Sand is related to the Loxton Sand of South Australia (Kenley, 1971). The Dorodong Sand also correlates to the Moorabool Viaduct Sand of the eastern Otway Basin.

The laterite horizons of the Dundas Plateau mentioned above are hosted within the Plio-Pleistocene Heterogeneous Aquifer in the west of the basin and are up to 10 m thick. They are stratigraphically positioned to imply that laterisation was dominant in the land-scape from mid to late Pliocene (Kenley, 1971; Wallace *et al.*, 2005). Similarly, the Timboon Surface is a ferricrete that covers the Plio-Pleistocene Heterogeneous Aquifer (a Pliocene sand) in the Colac region. Laterite and ferricrete development is explored further in Section 3.3.

The Whalers Bluff Formation was possibly deposited in a Pleistocene estuarine to littoral environment and comprises sandy limestone and bioclastic calcarenite with abundant oyster shells and a lenticular basal clay, also fossiliferous (Kenley, 1971). As the base of this unit constitutes an unconformity, it overlies a karst horizon in the Port Campbell Limestone, the Formation has a quite heterogeneous lithology, sometimes incorporating gravel and pebbles. Holdgate & Gallagher (2003) record the Whalers Bluff Formation as having a similar lithology, however, they describe a different depositional environment for it. Palaeoenvironment was deduced by them to be mid-shelf marine (100 m deep water) with deposition occurring in the middle Pliocene. Thick equivalent deposits are shown to exist as canyon fill in Miocene-cut canyons (Holdgate & Gallagher, 2003).

The Hanson Plain Sand comprises shelly sand and silt deposited in a near-shore regressive environment. It outcrops in the Port Campbell Embayment and is coeval with the Whalers Bluff Formation of the Gambier and Tyrendarra Embayments (Holdgate & Gallagher, 2003). It is considered equivalent to the Moorabool Viaduct Formation and was previously attributed to that formation.

The Bridgewater Formation is represented by Pleistocene strandlines composed of carbonate bioclastic sand (Kenley, 1971). These aeolian dunes formed sub-parallel to the coast, transverse to the prevailing wind direction (Sprigg, 1952). Holdgate & Gallagher (2003) described the Bridgewater formation as a Middle to Late Pleistocene aeolian sandstone, often cemented with carbonates.

The Malanganee Sand is aeolian in origin and consists of fine to medium grained quartz sand, probably derived from the Pleistocene dunes, the Bridgewater Formation (Kenley, 1971).

The Coomandook Formation comprises Pleistocene sandy limestone and calcarenite with minor skeletal fragments and lithics from older units and minor clay or silt (Tickell *et al.*, 1992).

Sub-aquifer Two

Sub-aquifer Two is composed of materials with lower hydraulic conductivity than that of Sub-aquifer One. The Padthaway Formation is one such unit; it forms the swales between the Bridgewater Formation dunes and was deposited in swampy and lagoonal environments. It has poor drainage characteristics, associated with its lithology of silts and clays with minor sand and carbonates. Sub-aquifer Two also comprises various unnamed lacustrine sediments, clay-rich alluvium deposits and some lunettes.

The Padthaway Formation is the main component of Sub-aquifer Two. It was deposited in a lacustrine environment, specifically, in the lagoons formed in the swales between the Pleistocene strand lines of the Bridgewater Formation (Sprigg, 1952). The Padthaway Formation comprises sediment of low hydraulic conductivity, such as silt, marl, sand, sandy clay and minor freshwater carbonates (Morand *et al.*, 2003). Its distribution is mostly limited to the Gambier Embayment.

Elsewhere, Sub-aquifer Two consists of Recent unnamed lacustrine deposits, for example, those associated with the previously larger extent of Lake Corangamite (Currey, 1964). The lake formed in response to basalt deposits blocking drainage lines (see Section 3.3 for more detail).

2.4 Basin Structure & Tectonic History

Of interest to the interpretation of the Otway Basin's hydrogeological flow system is the structure and tectonic evolution of its rocks. The geometry of the hydrostratigraphic units, shaped by deformations during tectonic uplift and subsidence, influences the physical flow regime of the basin. Additionally, observations of increased hydraulic gradients and of discharge coincide with tectonic faults, fractures and hydrostratigraphic boundaries (Wopfner & Douglas, 1971; Waterhouse, 1977; Love *et al.*, 1993).

2.4.1 Structural Setting

The Otway Basin formed in a continental margin setting, due to Jurassic-Cretaceous rifting between Australia and Antarctica, prior to which, these land masses were joined as part of the supercontinent Gondwana (VonDerBorch *et al.*, 1970). During rifting, the continuity of the basement structure was extensively disrupted by faulting. This syn-depositional faulting variably affected the younger rift-related passive margin sediments (Leonard *et al.*, 1983), leading to the development of hydraulically influential monoclines and structural highs.

The basin is subdivided into several distinct embayments, bounded by faults and basement highs. From east to west these are commonly, but not exclusively, known as the Gambier Embayment, the Tyrendarra Embayment, the Port Campbell Embayment and the Barwon Downs Graben or Embayment (Leslie, 1966; Wopfner & Douglas, 1971). These embayments are respectively separated by the basement highs: the Dartmoor Ridge, the Warrnambool High and the Barongarook High as shown in Figure 2.2. The Otway Ranges form the basin's southeastern margin and act to separate the Otway Basin proper (to its west) from the coeval Torquay Sub-Basin (to its east and northeast; not addressed in this thesis).

As previously discussed, the Otway Ranges are composed of the Eumeralla Formation, which is the main member of the Lower Cretaceous Basement. These rocks extend below the Tertiary and Late Cretaceous sediment of the basin, and crop out at the surface on the northern and eastern margins. In addition, the basement to the north consists of older igneous and metamorphosed rocks. The western most of these is the Padthaway Ridge, northwest of the Gambier Embayment. The ridge consists of Ordovician granites which outcrop at the surface in small areas and were deformed in the waning stages of the Delamerian Orogeny (Love, *et al.*, 1993). Eastwards, north of the Dartmoor Ridge, the Dundas Plateau consists of the Cretaceous Otway Group and Cambrian plutons and sediments that were heavily deformed by the same orogenic activity during the Cambrian and Ordovician (Miller *et al.*, 2002). Further east, north of the Tyrendarra Embayment, the Grampians Ranges are composed of Silurian sediments (Gibson & Nihill, 1992). The basement to the north of the Port Campbell and Barwon Downs Embayments comprises Cambrian and Ordovician marine sediments that were deformed during a younger orogenic event which resulted in the Lachlan Fold Belt. It is therefore plausible that the major structural geometry of the Otway Basin, which is spatially variable, resulted from the superimposed deformations associated with each of these orogenies (Basin evolution is discussed further in Section 2.4.2, below; Miller *et al.*, 2002)

The subsurface structure of the basin's margins was first explored by an early aeromagnetic survey conducted by the Geological Survey of South Australia. A basement ridge structure between the Murray and Otway Basins was recognised, and eventually identified as the Padthaway Ridge (Sprigg, 1952). The offshore extent of the thick layer of Tertiary sediments was first clearly outlined by the use of aeromagnetic surveys in 1960 and 1961. Also at this time, two gravity surveys, one from Victoria and one from South Australia, provided the first solid evidence of the onshore extent of Tertiary sediments and the depth to basement (Wopfner & Douglas, 1971).

As to the offshore extent of the basin sediments, thick sequences, sometimes more than 4 km thick above the Lower Cretaceous Basement, have been confirmed in petroleum and gas exploration wells positioned up to 50 km offshore. These are depicted in the cross sections of Figures 2.5 and 2.7. Even so, the wells furthest from the coast are on the continental shelf rather than the slope. However, the presence and thickness of the strata beyond the edge of the shelf have been confirmed by seismic survey. Such seismic surveys have also effectively illustrated distribution and intensity of faulting; faults are notably omitted by the cross sections of Figures 2.5 and 2.7, but the main faults are depicted in Figure 2.21. Four published interpreted seismic cross sections, which extend across the entire continental slope, are reproduced here for comparison in Figures 2.17, 2.18, 2.19 and 2.20. Locations of all the cross sections are shown in Figure 2.3. These seismic sections clearly show the distribution of Late Cretaceous and Tertiary sediments up to and beyond the toe of the continental slope. In can be seen that fault regime is dominated by south-dipping normal faults, typical of rifting and subsidence in sedimentary basins. In particular, Figure 2.17 shows the incision into, and consequent exposure of, the Lower Tertiary Sandy and Upper Cretaceous Aquifers by the sub-marine canyons on the continental slope.

Recent faulting has uplifted the Upper Tertiary Carbonate Aquifer in places, e.g. the upthrown northern block of the Kanawinka Fault, and exposed it at the surface (Kenley,



Figure 2.17. An offshore interpreted seismic cross section from Moore et al. (2000). Seismic line is number 137-03. Dark Green is the Lower Cretaceous Basement. Pale green contains the Upper Cretaceous Aquifer and the Upper Cretaceous Aquitard. Bright Yellow contains the Upper Tertiary and Lower Tertiary Aquitards and the Lower Tertiary Sandy Aquifer. Pale yellow contains the Upper Tertiary Aquitard, and the Upper Tertiary Carbonate, Pliocene-Quaternary Volcanic and Plio-Pleistocene Heterogeneous Aquifers.



Figure 2.18. An offshore interpreted seismic cross section from Geoscience Australia (2005). Seismic line is number 137-04. The shaded area is part of the Cretaceous Sherbrook Group, part of the Upper Cretaceous Aquifer.

1962). At other locations, as previously discussed, recent uplift has resulted in the erosion of parts of the Upper Tertiary Carbonate Aquifer. Faults and joints in the aquifer have not disjointed its continuity, rather they act as conduits for preferential flow (Love *et al.*, 1993). The Narrawaturk Marl, Mepunga Formation and Gellibrand Marl of the Upper Tertiary Aquitard have generally south-dipping beds with some gentle folding, axes oriented NE-SW, and host little faulting; this structural trend is also apparent in the Clifton Formation of the Upper Tertiary Carbonate Aquifer (Holdgate & Gallagher, 2003).

Despite pervasive block faulting within the Wangerrip Group, continuity and connectivity of its Formations, e.g. the Lower Tertiary Sandy Aquifer, are maintained (pers. comm. Schneider C., 2005). In contrast, faulting at deeper levels or in older formations, such as the Upper Cretaceous Aquifer, has produced larger displacements and has created viable gas and petroleum traps. The Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer have also been affected by folding. In the Port Campbell and Barwon Downs Embayments, the aquifers are deformed over the Ferguson Hill Anticline and Bar-







Pale green contains the Upper Cretaceous Aquifer and the Upper Cretaceous Aquitard. Bright Yellow contains the Upper Tertiary and Lower Tertiary Aquitards and the Lower Tertiary Sandy Aquifer. Pale yellow contains the Upper Tertiary Aquitard, and the Upper Tertiary Carbonate, Pliocene-Quaternary Volcanic and Plio-Pleistocene Heterogeneous Aquifers. Figure 2.20. An offshore interpreted seismic cross section from Moore et al. (2000). Seismic line is number 137-09. Dark Green is the Lower Cretaceous Basement.

won syncline (Figure 2.21). A more detailed description of folding and faulting in that region is provided by Medwell (1971).

The groundwater of the embayments, or sub-basins, has often been studied separately, mainly for the reason that the basement highs create hydraulic axes, limiting groundwater flows either to one embayment or the other (Leonard *et al.*, 1983; Laing, 1989; Love *et al.*, 1993). However, the structural analysis of the hydrostratigraphy shows that the major aquifers are laterally contiguous, indicating that hydrogeological similarities should exist between the sub-basins and some hydraulic connection exists.

Several authors have indicated the Otway Basin is comparable to other basins, having similar tectonic, structural, sedimentological and/or hydrochemical characteristics (Johns, 1968; Blake, 1989; Laing et al., 1989; Finlayson et al., 1996). In particular, the other major coastal basin of Victoria, the Gippsland Basin, provides some similarities in the areas of sedimentary sequence, groundwater-petroleum interaction and high groundwater temperatures (Nahm, 2002). Suitable international comparisons include the North Atlantic Coastal Plain (USA), East African margin basins, and the London Basin (UK).

2.4.2 Structural & Tectonic History

Sprigg (1952) outlined the basin's structural history, recognising three main episodes of deformation: pre-Tertiary uplift, intermittent Tertiary faulting and folding and regional warping in the Quaternary. In addition to these events, it is now accepted that rifting and extension were important in the basin's initiation (Hill *et al.*, 1995; Finlayson *et al.*, 1996; Miller *et al.*, 2002). The sequence of deformations is described below and summarised in Figure 2.22.

2.4.2.1 Onset of rifting

The initiation of the Otway, Gippsland and Bass Basins was by the rifting of East Gondwana separating the Australian and Antarctic plates in the late Jurassic. This extensional regime (Figure 2.22), accompanied by regional subsidence, continued throughout the Early Cretaceous until the final breakup of the two plates had occurred by about 95 Ma (Hill *et al.*, 1995; Figure 2.22). The sediments of the Otway Group were deposited and heavily faulted during this rifting and subsidence, with most major faults showing normal displacement (Figure 2.17).

The style of extension during rifting is somewhat contentious. Hill *et al.* (1995) have proposed N-S extension, dominated by E-W trending faults with subsidiary NW-SE and NE-SW fault trends due to pre-existing rock fabrics. However, other authors (e.g. Geary & Reid, 1998) support NNE-SSW or NNW-SSE oriented extension. As explained earlier,







Figure 2.22. Tectonic history of the Otway Basin. Note the coincidence of inversion events with unconformities in the sedimentary sequence. A—detachment from Antarctic plate; B—uplift of the Otway Ranges; C—intermittent volcanism; D—opening of the Southern Ocean and establishment of the circumpolar current; E—arc collision with New Guinea; F—intense, explosive volcanism; G—uplift leads to Plio-Pleistocene unconformity. Timing of events was taken from Hill *et al.* (1995); Price *et al.* (2003); McGowran *et al.* (2004) and Wallace *et al.* (2005).

the Palaeozoic basement rocks to the north of the Otway Basin are represented by the Delamerian Fold Belt in the west and the Lachlan Fold Belt in the east. The division between these belts, when extrapolated south into the Otway Basin, resides to the east of the Timboon Fault (Figure 2.21). It has been suggested by Miller et al (2002) that the eastern and western portions of the Otway Basin where influenced by the Lachlan and Delamerian Fold Belt geometries respectively during early rifting. This resulted in the Early Cretaceous faults having different strike and dip orientations on either side of the division; in the western zone NW-SE faults dipped both to the NE and SW whereas the NE-SW faults of the eastern zone generally dipped to the NW.

2.4.2.2 Pre-Tertiary inversion and ongoing extension

In response to the detachment of the Australian and Antarctic plates, an episode of rapid uplift, cooling and inversion followed the Cretaceous extension, lasting from 95 to 80 Ma (Hill *et al.*, 1995; Figure 2.22). This event was responsible for the uplift and deformation of the Otway Group which formed the Otway Ranges and the Dundas Plateau (Sprigg, 1952). Up to 2 or 3 km of rock and sediment was eroded from the Otway Group in some places, resulting in a major unconformity between it and the younger Sherbrook Group (Hill *et al.*, 1995; Figure 2.22). Northward drifting of the Australian plate began in the late Cretaceous, also in response to the final stage of continent breakup.

After this inversion, a second episode of rifting resumed and continued throughout the Late Cretaceous (Figure 2.22). During this extension, the Sherbrook Group was deposited and deformed by syn-depositional normal faulting. The extension regime had changed by this time to be aligned NE-SW, with faults developing in a NW-SE orientation, dipping to the SW (Miller *et al.*, 2002; Figure 2.21). Extension was accompanied by continued northward drifting of the Australian plate and thermal subsidence (Hill *et al.*, 1995). The importance of this period of the basin's development is reflected by the fact that the members of the Sherbrook Group, i.e. the Upper Cretaceous Aquifer and the Upper Cretaceous Aquitard, have much higher maximum thicknesses than younger units (Figure 2.4). However, due to the uplift in the vicinity of the Otway Ranges, the Sherbrook Group is thin or missing in that area (Figure 2.10).

2.4.2.3 Tertiary extension and intermittent inversion

Not long before the beginning of the Tertiary, a short-lived inversion event may have occurred, reactivating normal faults in with a reverse sense of movement. This event potentially coincides with the unconformity between the Sherbrook and Wangerrip Groups (Figure 2.22). It is thought to have been caused by intrusions (e.g. Figure 2.17) but is still poorly time-constrained (Hill *et al.*, 1995). Extrusive volcanism did not start until later, around the mid-Palaeocene (Figure 2.22).

After this short inversion, the early Tertiary was characterised by continued thermal subsidence and deposition of the Wangerrip Group (Figure 2.22). Concurrently, seafloor spreading aided the northward drift of the plate. Also during this time, intermittent volcanism began (Price *et al.*, 2003).

Extension resumed during the Eocene and caused more normal faulting. The deposition of the Nirranda Group occurred during this phase which is characterised by thermal subsidence. During the Eocene, extension and sea floor spreading resulted in the opening of the Southern Ocean (Geary & Reid, 1998). The rifting was interrupted by occasional inversion events throughout the Tertiary, one such event causing the unconformity above the Nirranda Group (Figure 2.22) and others possibly responsible for some further uplift and erosion of the Otway Ranges (Hill *et al.*, 1995).

2.4.2.4 Neotectonic events

An extensional regime persisted into the early Miocene, accompanied by ongoing seafloor spreading, northward drift, thermal subsidence and intermittent volcanism. Deposition of the Heytesbury Group began in the late Oligocene and continued into the late Miocene (Figure 2.22). However, extension and volcanism ceased before the mid Miocene (Price *et al.*, 2003).

Towards the end of the early Miocene, plate drift and seafloor spreading resulted in the arc collision of the northern margin of the Australian plate with Papua New Guinea, causing extensive uplift of northern Papua New Guinea (Hill *et al.*, 1995). Around the same time, at about 12 Ma, the stress regime at the Autralian plate's southeast margin with the Pacific Plate also became compressional, resulting in the onset of uplift of the Southern Alps, New Zealand (Dickinson *et al.*, 2002). The stress of these compressional regimes at two of the plate's margins not only resulted in proximal uplift, but was also transferred across the Australian plate and accommodated by reverse reactivation of existing normal faults in the southern margin basins, i.e. the Otway, Port Phillip and Gippsland Basins (Hill *et al.*, 1995; Dickinson *et al.*, 2002).

Erosion driven by this late Miocene inversion resulted in the Miocene-Pliocene unconformity, which is visible at numerous outcrops throughout the Otway Basin (Dickinson *et al.*, 2002).

Uplift and deformation continued after the Miocene-Pliocene unconformity (Figure 2.22). Evidence includes the tilting of the Pliocene and Pleistocene strandline systems. The preservation and apparent progradation of the dune systems was possible only because of tectonic uplift (Murray-Wallace *et al.*, 2001; Wallace *et al.*, 2005). The axes of this uplift



Figure 2.23. Elevation of Pliocene and Quaternary strandlines in the Otway and Murray Basins showing two major uplift axes from Wallace *et al.* (2005).

can be seen in Figure 2.23. In the northern Gambier Embayment, where the dunes trend NW-SE, the interdune swale lines dip to the NW by more than 0.01° (Sprigg, 1952). This northwards inclination of the once-horizontal shoreline features is in keeping with uplift along the ENE-WSW Gambier Axis in the south (Figure 2.23). The strandlines of the Tyrendarra Embayment lie south of the E-W oriented Western Highlands Axis and are oriented sub-parallel to it. Uplift has been detected in this region by the elevation of different dune ridges and Wallace *et al.* (2005) reported a difference of 240 m elevation over 80 km. This is a southwards tilt of 0.17° . This epeirogenic uplift is attributed, at least in part, to a diapiric effect associated with the contemporary intense volcanism of the Western Plains Province (Murray-Wallace *et al.*, 1999; Price *et al.*, 2003; Figure 2.22).

However, Sandiford (2003) linked Pleistocene deformation to the late Miocene inversion described above, which is associated with the development of a compressive stress regime over the Australian plate. Evidence of this connection is the displacement of Pliocene strandlines in the area of Cobden (location: Figure 1.2) over faults associated with the Ferguson Hill Anticline, Otway Ranges (Figure 2.21). This indicates reactivation of Cretaceous faults that played a role in the uplift of the Otway Ranges. It is possible that the two factors, i.e. the volcanism and the cross-plate compression, are combined in this persistent neotectonic event.

The Plio-Pleistocene uplift is most likely the cause of the dome structure apparent in Figures 2.13 and 2.15, which is described as the Tarpeena-Dartmoor Upwarp by Grimes (1994) and the Gambier Axis by Kenley (1971) and Love *et al.* (1993). The doming must be at least as young as post-Miocene, as it has resulted in the erosion of the upper parts of the Gambier Limestone. The southern margin of the dome coincides with the location of the older, south-dipping normal Tartwaup Fault (Miller *et al.*, 2002) and may be related to a reactivation of this structure. The deposition of the Pleistocene Bridgewater Formation over the region of eroded limestone indicates a hiatus in the uplift. However, the warping of the dune formations suggests the zone was reactivated later. This fits well with a history of waxing and waning volcanism over this time (Sprigg, 1952; Thompson, 1972; Price *et al.*, 2003).

Similar dome structures are not seen in the morphology of the hydrostratigraphic units elsewhere in the Otway Basin because of the position of the uplift axes. As can be seen in Figure 2.23, the Western Highlands Uplift Axis proposed by Wallace *et al.* (2005) is generally in the same orientation and position as the basin margins (c.f. Figure 2.9). However, in the region of the Tarpeena-Dartmoor Upwarp, the Gambier Axis encroaches seaward into the basin's sediment pile, in an orientation almost orthogonal to the basin margin.

2.5 Synthesis

The Otway Basin initiated on the southeastern plate margin of the Australian plate as a rift opened between this plate and the Antarctic plate. Ongoing rifting and subsidence during the Cretaceous and Tertiary allowed for episodes of widespread sedimentation. Syn-depositional normal faulting characterised periods of extension. Successive transgressions and regressions resulted in the deposition of clastic and carbonaceous marine and non-marine sediment sequences. The sedimentary groups are divided by regional unconformities caused by uplift and erosion during the inversion events that punctuate the extentional regime. Recent volcanism deposited basalt over much of the current surface of the basin and is probably related to neotectonic events.

The structure of the basin, i.e. its basement and the aquifers and aquitards of its strata, is a product of these processes. Structure contour maps and isopach maps created in this study confirm current knowledge, including the basement highs that separate the sub-basins or embayments of the basin and cause thinning of otherwise contiguous layers. Generally most of the hydrostratigraphic units are thinner at the basin margins and become thicker towards the ocean. Most of the aquifers and aquitards continue beyond the coastline, some way into the continental shelf, or even to the edge of the continental slope. Hydrostratigraphic units of the basin are delineated according to geological characteristics and are summarised briefly below.

The Lower Cretaceous Basement is arkosic siltstone and mudstone, although in some areas the basin strata overlie older bedrock, composed of metamorphic or plutonic rocks. The Upper Cretaceous Aquitard and the Lower Tertiary Aquitard are mostly deep marine mudstone with some carbonaceous clay and glauconite. The Upper Cretaceous Aquifer has upper and lower parts that are connected in places and separated by an aquitard in others. Both parts have variable grainsize and environments of deposition; they are dominated by quartz sand/sandstone and gravel with minor siltstone and carbonate minerals. The Lower Tertiary Sandy Aquifer has a similar lithology, with minor clay. The Upper Tertiary Aquitard and the Upper Tertiary Carbonate Aquifer are shallow and deep marine strata of carbonaceous clay, silt, mudstone and limestone with minor quartz clasts and some bioclastic layers.

The Pliocene-Quaternary Volcanic Aquifer is composed of basalt flows, scoria cones, tuff and stony rise deposits. The Plio-Pleistocene Heterogeneous Aquifer is a complex group of stratigraphic units, mainly comprising siliciclastic Pliocene sand and Pleistocene calcarenite and alluvium. These two youngest of the aquifers cover the vast majority of the surface of the Otway Basin and are responsible for the numerous karst and volcanic landforms present.

The Otway Basin is an incredibly diverse geological province which hosts valuable natural resources, onshore and offshore. The region has been subject to seismic and drilling exploration for hydrocarbons, rocks such as scoria and basalt are quarried and the groundwater resources of the aquifers are widely used.

Chapter 3

Climate, Eustasy and Hydrologic Balance

3.1 Introduction

This Chapter provides an analysis the geological history of the area with the focus on the development of a hydrologic history based on past environments and climates. Importantly, the palaeoclimate is compared to what is known about the current climate and atmospheric circulation patterns. Accordingly, both ancient and modern climate change events and/or processes are discussed. Therefore, the use of the term *climate change* in this study can denote natural fluctuations in climate as well as anthropogenic climate change.

Reconstruction of the palaeoenvironments at various times requires evidence from sedimentology, geochronology, palynology, biostratigraphy, palaeontology, etc. and these bodies of literature are drawn upon to create a timeline useful to this study. In addition, numerical palaeoclimate modelling literature is incorporated. Throughout, modern climate mechanisms and their changes within historical times are used to inform interpretations of the literature.

The main focus in this chapter is on the current climate and the climatological history of a time frame commensurate with the residence time of some of the oldest water in the Otway Basin. One estimate of residence time is available from a 14 C age of groundwater from the Clifton Formation, situated about 50 km from the coast in the Tyrendarra Subbasin, found to be just over 20,000 radiocarbon years (Bennetts, 2005). Therefore, the last 20,000 years or so is the minimum relevant time frame for this study. However, in order to encompass a complete glacial cycle, the time period from the current interglacial to the last interglacial, at about 130 ka, was adopted instead. The current climate of the

region is discussed first in Section 3.2 and is followed by the palaeoclimate reconstruction in Section 3.3.

3.1.1 Atmospheric and oceanic circulation

A generalised atmospheric circulation model of the Earth is represented in Figure 3.1. These patterns, along with the position of the continents and thermohaline variations are also the driving force behind oceanic circulation (Woodroffe, 2002). The climate of Australia is of course influenced by both oceanic and atmospheric circulation, but the main topic of concern in this study is the changes in delivery of precipitation to the southeast of the continent, which are most sensitive to changes in atmospheric circulation.

The climate of southeast Australia is controlled partly by the Ferrel cell circulation. The westerlies at approximately 30°S latitude, north of the polar front, are associated with the ascending arms of the Ferrel and Polar cells. The ascending air creates low pressure systems at the surface; these westerly lows deliver rain to the Otway Basin region in the winter. Further north, the descending arms of the Ferrel and Hadley cells result in a string of high pressure systems known as the sub-tropical anticyclonic belt (STA). Currently, the STA resides at the latitudes of approximately $\pm 30-35^{\circ}$, north of the Otway Basin, where it perpetuates relatively dry climates (Wells, 1997). At lower latitudes, the intertropical convergence zone is more humid by comparison. The latitudes of these low- and high-pressure systems can shift, resulting in long term climate change. Indeed, the northwards and southwards migration of the STA in the past is thought to be responsible for some aspects of climate change (Harrison, 1993; Ayliffe *et al.*, 1998). In addition, seasonal shifts result in the seasonal climate variations of southeast Australia (Sturman, 1996).

In addition to meridional circulation, the zonal atmospheric system of the Walker Circulation is an important influence on oceanic circulation and, in turn, precipitation in Australia. The Walker Circulation is usually quantified by the southern oscillation index (SOI), a measure of the difference of the pressure at Darwin and Tahiti. When the SOI is close to zero, the typical Walker Circulation ensures the easterly movement of the trade winds across the Pacific Ocean at low latitudes (Figure 3.2; top). The easterlies push warm water at the surface of the ocean to the west and force upwelling of cold water near the coast of South America (Figure 3.2; top). The warm ocean and ascending air in the west is associated with the monsoon in the north Australia. When the circulation is weak, the SOI drops below zero and equatorial winds and oceanic patterns are reversed (Figure 3.2; bottom). This constitutes an El Niño event and reduces rainfall over northern and eastern Australia (Woodroffe, 2002). Due to the connection of meridional cells, these circulation and precipitation patterns in the north have an effect on the position of the STA to the south, and can therefore influence the climate of the Otway Basin.



Figure 3.1. Vertical and meridional atmospheric circulation: Hadley, Ferrel and polar cells. Modified from online source http://rst.gsfc.nasa.gov/Sect14/Sect14_1c.html. Note the westerlies at approximately 30°S, north of the southern polar front. ITCZ—Intertropical Convergence Zone.

3.1.2 Climatological parameters

The two most important climatological parameters for groundwater studies are evaporation and precipitation because together they influence how much deep drainage may occur. The term deep drainage is synonymous with recharge and describes the process of soil water becoming groundwater, this occurs by vertical drainage of water through the unsaturated zone towards the watertable.

Evapotranspiration can be expressed as actual or potential and is affected by vegetation cover and type, soil type, depth to the watertable or availability of water and climate (Chiew *et al.*, 2002). Actual or potential evapotranspiration can be calculated using the Penman-Monteith equation, which requires inputs of temperature, wind speed, insolation and humidity. Satellite data can be used to perform this calculation (Choudhury, 1997). Actual evapotranspiration can also be measured by field lysimeters (e.g. Bethune & Wang, 2004;) or via remote sensing by constructing an energy balance using infrared light or microwaves (e.g. Bastiaanssen *et al.*, 1998a & 1998b).

Evapotranspiration must be used in conjunction with precipitation in order to judge the expected deep drainage or recharge from climatic conditions. The precipitation/evapotranspiration ratio (P/E) and the precipitation/potential evapotranspiration ratio (P/PE) are the parameters used; they are defined by the following equations:



Walker Circulation during an El Niño



Figure 3.2. A diagram showing the atmospheric circulation patterns at the surface and aloft during normal Walker circulation (top; $SOI \approx 0$) and El Niño conditions (bottom; SOI < 0). After Bureau of Meteorology (2005b).

$$P/E \ ratio = \frac{annual \ average \ precipitation \ (mm)}{annual \ average \ actual \ evapotranspiration \ (mm)}$$
(3.1)

$$P/PE \ ratio = \frac{annual \ average \ precipitation \ (mm)}{annual \ average \ potential \ evapotranspiration \ (mm)}$$
(3.2)

Most of the ratios discussed in this study are annual averages, however, the P/E ratio changes from month to month with seasonal shifts, especially in a Mediterranean-type climate such as that of the Otway Basin (Section 3.2). In the Otway Basin, the P/E ratio is high in winter and low in summer.

Climates with a P/PE ratio value greater than 0.65 are considered humid; those with a value of 0.50 or less are semi-arid (Williams & Balling, 1996). P/E and P/PE ratios for an area or time are generally related, although not equal or always proportional. If the fluctuations in the P/PE ratio of the past are known, it can be assumed that similar trends occurred in the fluctuation of the P/E ratio.

Quantitative estimates of the palaeoclimate P/E ratio for the Otway Basin have been derived from lake evaporation in comparison to precipitation (Jones *et al.*, 1998; 2001). The particular method for estimation of lake evaporation used was that for Complimentary Relationship Lake Evaporation (CRLE)(Morton, 1986) which takes into account the salinity of the lake water and has no need for estimations of the wind speeds of palaeoclimates. As lake evaporation is similar to potential evaporation, these values are interpreted to be P/PE ratios.

3.1.3 Environmental indicators of hydrologic balance

3.1.3.1 Glaciation

The climate history focuses largely on glacio-eustatic cycles as they are the prominent feature of the Quaternary palaeoclimate of any region on the globe and they are crucial for understanding several aspects of groundwater history. Inclusion of eustatic timelines is important because the aquifers of the Otway Basin are hydraulically connected to the ocean.

Sea levels can oscillate due to changes in global temperate and/or changes in the cryosphere. However, isostacy, associated with plate tectonics, mantle influences and glacial loading, also controls relative sea level (Lambeck, 2002). Notwithstanding the causes of sea level change, the influences it has on coastal groundwater systems include: change in pressure head in aquifers hydraulically connected to the ocean; change in position of the saline-fresh water interface; and change in continentality, resulting in reduction or increase of rainfall.

The other aspect of glaciation cycles that affects groundwater is the change to the global climate and atmospheric circulation. Changes to global temperature and hydrologic balance leads to a change in P/E ratio, which affects recharge. Also, global temperature change can lead to changes in sea surface temperature and therefore ocean and atomsphere circulation. This in turn can change climate patterns, again affecting the P/E ratio or its distribution pattern. Too often it is assumed that interglacial periods are warm and therefore more humid, and that glacial periods are dry. This is the case in some regions (e.g. Preusser *et al.*, 2002). However, Nott and Price (1994) found that currently, during an interglacial period, the north Australian monsoon is weaker than it was during the last glacial maximum. There are many other cases which confirm that unravelling palaeo-climate requires more than correlation of events with global eustasy and glaciation; that oceanic circulation and associated atmospheric circulation is very important in dictating climate variations, both temporally and spatially (e.g. Hu *et al.*, 2003).

A convenient proxy of glacio-eustatic cycles is the isotopic composition of oxygen atoms of marine water, which is measured as the δ^{18} O value of minerals in marine rocks. The foraminifera, coral and other fossils of marine sediment are analysed for their oxygen isotope composition and from this, the composition of the ocean in previous climates can be deduced (e.g. Shackleton *et al.*, 1995; Shackleton 2000 & Waelbroeck *et al.* 2002). Greater δ^{18} O values for ocean water correlate to a greater proportion of heavy water molecules in the ocean, mainly due to the accumulation of light molecules in ice caps. Hence, the timing of glacial and interglacial cycles can be constrained using the isotopic ocean composition. Glaciation results in "heavier" oceans, whereas interglacial periods are marked by a lighter oceanic composition.

The detailed chronology of marine isotope composition has resulted in the definition of marine oxygen isotope stages, identified by MIS numbers. Each stage relates to a particular glaciation or deglaciation, or even stadial and interstadial events. The concurrent fluctuations of relative sea level and marine oxygen isotopes are depicted in Figure 3.7. MIS numbers are listed in this figure and are also used in Figure 3.10. The description of the integrated palaeoclimate history for the Quaternary is divided into temporal stages with reference to both the relative dates of the MIS numbers and absolute dates from various methodologies.

3.1.3.2 Lake level and lake salinity

Lake environments are useful tools in palaeoclimate research. Lacustrine sediment geochemistry and geochronology can be used to reconstruct depth of water and its salinity in the past. Similarly, biological indicators such as polen, foraminifera or diatoms can also be used to infer the past conditions (e.g. Dodson, 1974; Moos *et al.*, 2009). Once factors of lake level (from water depth) and salinity have been thus ascertained, a lake balance can be constructed which is used to model the past fluctuations of precipitation and evaporation. This method is particularly useful for closed-basin lakes whose level and salinity are strongly reliant on the balance of precipitation and evaporation. In simplified version, the lake balance can be expressed as:

$$lake \ volume = inflow_{[P]} + inflow_{[R]} - outflow_{[E]}$$

$$(3.3)$$

where $inflow_{[P]}$ is the precipitation incident on the area of the lake's surface, $inflow_{[R]}$ is the surface runoff from surrounding catchment proportional to precipitation and $outflow_{[E]}$ is loss of water through evaporation from the lake surface (Bowler, 1981).

Use of physical lake balances to illustrate climatological fluctuations in P and E is not relevant for lakes that are strongly controlled by groundwater (Bowler, 1981), as additional terms of groundwater inflow and outflow would have to be added to the above equation. Lake Keilambete and other crater lakes of the Otway Basin are presented as being reliant on precipitation and surface drainage (e.g. Bowler, 1981; Jones *et al.*, 2001), as presumably the 6 m thick clay layer of the flat lake bed of Lake Keilambete (Bowler, 1981) is considered a barrier to groundwater flow. However, lake levels of the western Victorian volcanic lakes have long been recognised as expressions of the level of the watertable (Maddocks, 1967; Gill, 1989) and detailed hydrogeological investigations of the lakes find they are indeed not closed systems (Coram, 1996; Adler 2003; Adler & Lawrence, 2008).

An alternative but similar approach to lacustrine palaeoclimate takes into account the inflow (baseflow) and outflow (recharge to groundwater) of groundwater to and from the lake as a contributing factor to lake levels and salinity (e.g. Shapley *et al.*, 2009). Even though some studies of lakes do take groundwater into account, variability with time in the parameters is often not incorporated and sometimes only inflow can be considered (e.g. Jones *et al.* 2001). For example, Ahlberg *et al.* (2001) acknowledged groundwater inflow as a factor that can affect lake geochemistry but disregarded its influence in their final conclusions in order to avoid complexity beyond logical resolution.

Overall, the results of palaeoclimate studies on lakes in the Otway Basin as closed systems are valid because the volumetrically important terms of the lake budgets are precipitation and evaporation, rather than groundwater flux (Coram, 1996). In further support, modern observations show that lake depths and salinity levels are strongly affected by changes in surface influx (Timms, 2005).

3.1.3.3 Lunette formation

Lunettes are generally divided into two groups: sand lunettes and clay lunettes (Stephens & Crocker, 1946), and those of the Otway Basin are mostly clay dominated. Bowler (1983) presented a theory of clay lunette formation that requires wet and dry cycles of the lake to allow clay pellet formation on exposed lake beds and transportation to the lunette. Once deposited, the hygroscopic nature of the clay pellets, due to their salt content, enables the pellets to absorb moisture from the atmosphere and adhere to neighbouring pellets, creating a cohesive layer that becomes harder with drying. Thus, the location and timing of lunette formation can be informative about lake conditions in the past, relating to both lake level and salinity. As many of the lakes in the Otway Basin are affected by some form of groundwater-surface water interaction, past lake conditions are very pertinent to this groundwater study.

3.1.3.4 Speleogenesis and speleothem development

The formation of solutional caves (speleogenesis) is related to percolation of rain below the soil zone and it occurs via various different physical and chemical processes. The morphology of the caves is usually indicative of the specific speleogenetic processes. Therefore, the study of the timing and mode of cave formation provides much information about the hydrological balance in the past, including the elevation of the watertable and the relative humidity of the climate.

Speleothem growth is the subsurface formation of precipitates due to flowing or dripping groundwater entering caves. Therefore, the formation of speleothems requires infiltration of water below the soil zone, i.e. a climate where effective precipitation levels maintain deep drainage. Speleothems often have discernible growth lines or horizons which can be correlated to absolute or relative dates. This provides a climate history for the region, annotated with humid or pluvial phases of speleothem development and arid/semi-arid periods which inhibited or precluded speleothem development.

3.1.3.5 Dune building

Barrier dunes of the Otway Basin are formed from both siliciclastic (Pliocene) and calcarenite (Pleistocene) sand. Although siliceous and carbonate dunes have their differences, the mostly aeolian nature of their deposition hints at several environmental factors prevailing during the time of their development. The main point is that extended periods of relatively arid climate are required for such pervasive aeolian deflation of sediment to build such dunes (Murray-Wallace *et al.*, 1999; Preusser *et al*, 2002). However, within dunes, palaeosol horizons, such as ferricrete and calcrete layers, may be found and can provide further information on palaeoclimate primarily because differing levels of humidity are required to form environments conducive to their formation.

3.2 Current Climate

The Otway Basin, like much of southeast Australia, has a highly seasonal climate with hot, dry summers and cool, wet winters. However, the proximity of the Southern Ocean acts to moderate temperature maxima and minima to a certain extent. As most of the rain falls in the colder months of the year, the climate is termed Mediterranean-type.

Typical wind patterns in the region, over the Southern Ocean and over the onshore portion of the Otway Basin, are south-westerly in summer, and in winter, westerly (Sturman, 1996). These patterns are dictated by latitudinal shifts in the global-scale atmosheric circulation. Specifically, the westerlies situated to the north of the polar front in Figure 3.1, associated with the ascending arms of the Polar and Ferrel cells, are the winds of the winter trend. The south-westerly winds further to the north are the summer circulation pattern. As the westerlies are associated with low pressure systems, they are often rain-delivering winds. However, the south-westerlies are part of the subtropical anticyclonic belt (STA) and are therefore drier.

This shift in circulations patterns is due to the fact that the major driving force of the system is the convection at the Intertropical Convergence Zone (ITCZ), which is caused by direct insolation near the equator. At different times of the year, as the location of the most direct insolation shifts north or south, the ITCZ shifts also (Sturman, 1996). With this shift comes a corresponding shift in the other meridional circulation systems, also affecting the position of the STA.

To illustrate the seasonal fluctuations, some monthly average records for observation sites within the Otway Basin are provided in Figure 3.3 (stations marked in Figures 3.4 and 3.5). These were chosen to represent different conditions that can affect climate locally. For example, Forrest lies at a higher altitude than the other sites as it is situated in the Otway Ranges, as a result it has the highest rainfall and the lowest winter temperatures. Warrnambool and Mt Gambier are near the coast, but at different latitudes. Warrnambool is the closest site to the coast and therefore exhibits the least temperature fluctuation throughout the year. Similarly, Mortlake and Naracoorte lie at different latitudes but both represent inland conditions.

The Otway Basin region receives an average of at least 600 mm of rain per year (Bureau of Meteorology data 1961–1990; 2005a). The trends in precipitation distribution are illustrated in Figure 3.4. The orographic effect concentrates rain strongly over the Otway Ranges in the south east; a slight increase in rainfall is also observable near the cliffs of Cape Bridgewater.

The annual actual evapotranspiration in the basin ranges from 400 mm to 700 mm (Figure 3.5). Average annual Class A pan evaporation can reach up to 1450 mm in some regions of the basin (Walker *et al*, 1990), much higher than the actual evaporation. Pan evaporation represents potential evaporation and the difference between it and actual evaporation proves that the hot summer months are also very dry. Despite the dry summers, the P/E ratio is greater than 0.5 for the remaining 9 months of the year (Sprigg, 1952).

To illuminate the recent temporal trends of precipitation and evaporation, several time series of 11-year averages of rainfall and temperature are plotted (Figure 3.6) for four locations in the basin (Figures 3.4 and 3.5) and a fifth: Melbourne (refer to Figure 1.1). Plot **a** shows annual, winter and summer rainfall for the first four locations (Figure 3.6). Prior to 1930 rainfall trends are fairly even; during the following two decades they are rising. Following this, the 1950s are characterised by falling precipitation at all sites



Figure 3.3. A time series of average monthly maximum temperature and average monthly rainfall at several sites in the Otway Basin. Plotted using Bureau of Meteorology data (2009a). These averages are constructed from the existing instrumental records at each site, which consist of differing time periods of variable length. Therefore the comparative relationships are indicative only.

except Naracoorte, where the decline is short-lived. The rainfall levels recover or fluctuate slightly and are sharply increased during the 1980s. A decline is recorded in the 1990s and beyond. Generally, the summer and winter rainfall trends mimic the annual trends and there are no notable long-term differences in the rate of rise or fall in the summer and winter records.

Minimum temperature time series for the same locations (Figure 3.6b) show that annual, summer and winter trends appear similar to each other. They are generally stable or rising, although a decline at Camperdown was observed between 1910 and 1920. However, the continuity and longevity of these records is not as good as those for rainfall. Therefore, the temperature records of Melbourne, being long and continuous and being for a location of comparable climate, are employed as a proxy.

Both minimum and maximum mean temperatures for Melbourne show increasing trends, regardless whether the data represents the mean annual, summer or winter temperature (Figure 3.6c). The rate of increase for each dataset was calculated from a line of best fit, along with R^2 values for the linear regressions (Table 3.1). The minimum mean











Figure 3.6. Time series of rainfall (a) and temperature (b & c). a) 11-year averages of total annual rainfall, total winter rainfall and total summer rainfall at four locations in the Otway Basin. Averages were calculated from an 11 year period and plotted for the median year; e.g. data point for the year 1960 is an average of the data from years 1955–1965. b) 11-year averages of annual average monthly mean minimum temperature, average summer monthly mean minimum temperature, average summer monthly mean sthe minimum temperature data presented in (b) but for Melbourne and also showing Melbourne mean maximum temperature. Calculated from data supplied online by the Bureau of Meteorology (2009b).

Data type	Time period	Rate of increase (°C/100yrs)	R^2
mean maximum	summer	0.53	0.40
	annual	0.47	0.44
	winter	0.59	0.39
mean minimum	summer	1.76	0.78
	annual	1.46	0.77
	winter	1.22	0.74

Table 3.1. Rates of increase in temperature for Melbourne calculated from lines of best fit for the data depicted in Figure 3.6c.

temperature has risen over the last 150 years more quickly than the maximum temperature. Some of this warming is likely to be due to the urbanisation of the city. However, the warming trend is confirmed up to a rate of $1^{\circ}C/100$ yrs for the Otway Basin region by integrated analysis of temperature data for Australia (Torok & Nicholls, 1996).

The combined impact of the fluctuating trends of precipitation and the steady increase in temperature is a changeable but decreasing P/E ratio. Warming trends for southeast Australia are expected to continue into the future, accompanied by lower rainfall (IPCC, 2007).

3.3 Palaeoclimate & Eustatic Fluctuations

3.3.1 Cretaceous & Early Tertiary

In the Early Cretaceous, the Australian plate was positioned next to Antarctica, at much higher latitudes than it currently resides. The southern margin of the plate, the region of the Otway Basin, was close to a latitude of 85° South, within the Antarctic Circle. It therefore probably experienced an annual period of extended darkness (Mehin, 1999). However, the mean global temperature, including temperature at high latitudes, during the Late Jurassic and Early Cretaceous was higher than those of today (Sellwood & Valdes, 2006).

Towards the Late Cretaceous, a decline in global temperatures saw the formation of a South Pole icecap, while Australia maintained a warm-temperate to subtropical climate (Mehin, 1999). The reason for this lies in the rifting of the two plates and sea floor spreading between them. This process allowed (a) the Australian plate to drift northward to warmer latitudes and (b) the Southern Ocean to circulate more freely. The circulation increased to a point in the late Eocene where it created the Antarctic Circumpolar Current, which climatologically isolated Antarctica (McGowran *et al.*, 2004). These different climatic conditions for the two continents continued and further changes occurred in throughout the Tertiary. However, the late Tertiary and the Quaternary are more relevant to the groundwater history, therefore, they are the focus of the following sections.

3.3.2 Pliocene (5.3 Ma - 2.6 Ma)

The Pliocene global climate fluctuated between glacial and interglacial periods, although the fluctuations were more moderate than Recent ones. Over the epoch, a long-term cooling trend was coincident with these dynamic changes. The climate of southeastern Australia migrated from warm-wet to cool-dry (Gallagher *et al.* 2003). Gallagher *et al.* (2003) deduced from foraminiferal and palaeobotanical evidence that the conditions of the Early Pliocene (c. 5 Ma) were 2–4°C warmer, with 50–70% more precipitation than today; and by comparison, the Early Pleistocene (c. 2 Ma) climate was only 0–2°C warmer, with up to 30% more precipitation.

The glacio-eustatic oscillations and the overprinting cooling trend of the Pliocene are both visible in the oceanic δ^{18} O record. (Figure 3.7, bottom). The glacial cycles become more intense towards the end of the epoch.

Ferricrete or laterite horizons (e.g. the Dundas and Timboon Surfaces; Section 2.3.9) are almost ubiquitous within Pliocene sediment of the Otway Basin. Laterisation is a process related to the stable position of the watertable and is often dominant in tropical climates. It follows, then, that extensive ferricrete development during the Pliocene may denote a wetter and warmer climate than today (Wallace *et al.*, 2005), which colludes with other climate proxies.

Another indication of the humid Pliocene climate for the basin is the onset of speleogenesis. The Naracoorte caves, hosted within the Upper Tertiary Carbonate Aquifer, began to form in the late Pliocene, although the main speleogenesis occurred in the Pleistocene, after 1.1 Ma (White, 1994).

Also at this time, lake formation occurred because of the eruption of the first phase of the Western District Province Volcanics, which consisted predominantly of thick (30–60 m) sheets of basalt that filled pre-existing channels (Currey, 1964; Coram, 1996). Basalt flows blocked and diverted the surface drainage of the area (Ollier, 1988) and, aided by the high P/E ratio, allowed the development of lakes. Evidence linking these lacustrine environments to the volcanics includes the geomorphology of today's remnant lakes. The characteristic flat bottoms of many of the shallow lakes in the Corangamite region indicate that tectonic subsidence was not a major factor in forming the depressions of lakes, rather the tops and edges of the basalt flows form the lake bottoms and banks respectively, further evidenced by the irregular shorelines of these lakes (Timms, 1992; Coram, 1996). Thus, development of lakes in the region was likely to have started during the Late Pliocene, or



Figure 3.7. Oceanic δ^{18} O levels as a proxy for global temperatures and glaciation after Waelbrocck *et al.* (top; 2002) and Wallace *et al.* (bottom; 2005). Vertically-oriented numbers (top) are MIS stages; RSL—relative sea level; ka—thousands of years ago; Ma—millions of years ago. Greater positive δ^{18} O values for ocean water correlate to "heavy" ocean and cold or glacial climate.

at least during the beginning of the Pleistocene, by which time the first phase had all but ended (Thompson, 1972).

3.3.3 Early and Mid Pleistocene (2.6 Ma – 130 ka; MIS 6 and older)

The climate of the Otway Basin in the Pleistocene can be summarised as fluctuating between arid or semi-arid glacial and interglacial stages with relatively humid phases interspersed during times of moderate sea level.

Throughout the Quaternary, glacio-eustatic oscillations intensified from their state in the Pliocene. This is indicated by the oceanic isotope composition (Figure 3.7). These successive warmer and cooler fluctuations, having a wavelength of roughly 100,000 years, were superimposed over a gentle cooling trend of a longer wavelength, lasting the whole of the Quaternary. The glacial and interglacial maxima are intermittently punctuated by stadials and interstadials.

After the relative tectonic quiescence of the Early- and Mid-Pliocene, the Late Pliocene and Early Pleistocene saw some tectonic activation as evidenced by uplift of the Pliocene strandlines (Wallace *et al.*, 2005; Figure 3.7; Section 2.4.2.4). The uplift and glaciation were together responsible for the formation and preservation of the Pleistocene strandlines of the Bridgewater Formation. These aeolian dunes have been shown, via various dating methods, to be younger with each successive range towards the coast and to coincide with sea level high stands, i.e. interglacial maxima or very warm interstadials (Huntley *et al.*, 1993; Murray-Wallace *et al.*, 2001). The correlation of specific continuous dune ranges in the Gambier Sub-basin with peaks in sea level and maximum depletion of oxygen-18 in the ocean is depicted in Figure 3.8; the dunes' locations are shown in Figure 3.9. Using an average, constant rate of uplift of 0.07 mm per year for the Quaternary, Murray Wallace *et al.* (2001) concluded that most of the sea level high stands recorded by the dunes were within 6 m of today's sea level (supported by Belperio *et al.*, 1995).

The dominance of the Bridgewater Formation dunes throughout the Pleistocene indicates a dominance of the aeolian processes of their formation. Sprigg (1952) documented many aeolian sedimentary structures of the dunes, including cross bedding and landwarddipping laminations. This widespread aeolian activity represents one or many periods of arid climate during the Pleistocene. Further evidence of aridity is the aeolian deflation of the Bridgewater Formation during the Late Pleistocene to form the Malanganee Sand (Kenley, 1971). Murray-Wallace *et al.* (1999) investigated in detail the facies relationships within the Woakwine Range (location: Figure 3.9) and confirmed its aeolian nature. Further, they attest to the abundant cool-water carbonate productivity on a wide continental shelf that would have been necessary for the development of the calcareous Woakwine dune range. This carbonate productivity prevailed due to the ideal conditions, namely,



Figure 3.8. Age distribution of high stand dune ranges of the Gambier Sub-basin correlated with ocean δ^{18} O after Murray-Wallace *et al.* (2001). Dune locations can be found by the names of the ranges mapped in Figure 3.9. A change towards a more negative value of δ^{18} O corresponds to a transgression and *vice versa* for regressions.

a lack of inhibiting clastic terrigenous input. The paucity of terrigenous sediment was due to the arid climate and the geomorphology of older dune ranges, which counteracted coastward drainage.

The convention is to associate arid climates with glacial maxima, when much of the hydrosphere is locked up in the glacial ice. For example, inland aeolian dunes of the Arabian Peninsula were found to represent sea level low stands, when exposed sediments were easily deflated (Preusser *et al.*, 2002). Nonetheless, the periods of aridity represented by the pervasive aeolian activity in the Otway Basin must coincide with the main dune-building episodes, namely, the majority of the Pleistocene interglacials and some of the hot interstadials (Huntley *et al.*, 1993; Murray-Wallace *et al.*, 2001; Figure 3.8).



Figure 3.9. Locations and names of Pleistocene dune ranges from Murray-Wallace et al. (2001).
However, the glacial maxima are also likely to have been arid times in the Otway Basin. Ayliffe *et al.* (1998) dated the speleothems of the Naracoorte caves using the 230 Th/ 234 U method and correlated relatively wet and dry periods with sea level and sea surface temperature fluctuations over the last 500,000 years. They found that both interglacial and glacial maxima precluded speleothem growth and were therefore relatively arid periods in the Naracoorte region. Interglacial maxima and the hottest interstadials were arid due to high temperatures causing potential evaporation to exceed actual evaporation; glacial maxima were arid due to reduced precipitation as a result of the changed hydrologic cycle, mentioned above (Ayliffe *et al.*, 1998).

Additionally, it appears that the arid glacial and interglacial maxima were punctuated by brief pluvial phases, possibly associated with the moderate conditions of stadials and cool interstadials. Evidence of the pluvials includes: lacustrine environments, calcarenite caps on the dunes, lunettes, speleogenesis and speleothems; each discussed below.

While the Bridgewater Formation was being deposited in the west, the centre and east of the Otway Basin was undergoing changes from extensive volcanism. Currey (1964) and Thompson (1972) propose that, based on the distribution of Pleistocene lacustrine sediments, by the completion of the first volcanic phase (around 1.8 Ma; Price *et al.*, 2003), the Corangamite region was fully submerged under an enlarged Lake Corangamite. To sustain the large lacustrine system, high precipitation and low evaporation rates would have been necessary; therefore, during periods of the mid- to late- Pleistocene, the region had a wetter climate than today with a high P/E or P/PE ratio.

Development of the calcretes which cap Pleistocene dune sequences is the main reason for dune preservation and fixation and is also a key to palaeoclimate evolution (Sprigg, 1952). Globally, and in the case of the Otway Basin, calcrete growth, especially in several generations intercalated with aeolian dune sediment, is linked to pedogenesis and represents a semi-arid climate with highly seasonal rainfall or other cyclical wet-and-dry periods (Spring, 1952; Cooke, 1975; Belperio *et al.*, 1995; Murray-Wallace *et al.*; Preusser *et al.*, 2002). Belperio *et al.* (1995) link calcrete formation to pedogenesis and palaeosol development, where calcrete caps often develop in the vadose zone on dune ridge tops during times of relative dune stability and plant colonisation. The role of vegetation is further evidence of the presence of effective precipitation which could sustain plant communities at the time of, or prior to, calcrete precipitation (Preusser *et al.*, 2002). Despite calcrete and carbonaceous soil development possibly denoting pluvial phases within the Pleistocene, overall, they denote a drier climate than that of the Pliocene, when ferricrete and iron-rich soil development dominated (Wallace *et al.*, 2005).

Unfortunately, calcrete layers are chronologically complex and often diachronous, and, although multiple generations are distinguishable, it is impossible to achieve stratigraphic correlation between different sites of the Gambier sub-basin (Belperio *et al.*, 1995). In addition, absolute dating of calcareous pedogenesis is problematic due to sometimes multiple episodes of recrystallisation (Callen *et al.*, 1983). This means that it is difficult to constrain the timing of the individual pluvial events responsible for the calcrete development.

In the late Pleistocene the palaeo-lake of Lake Corangamite declined from its peak indundation. The large lake was replaced by many smaller ones, mostly the same lakes which persist today, many of which exhibit lunettes. Lunettes were also deposited at various stages of the palaeo-lake's retreat (Currey, 1964). The lake regression is not well constrained in time and could be linked to a number of hydrologic or climatic causes. However, the Quaternary lunettes do provide insight into the palaeoclimate of the Pleistocene. Bowler's theory concerning clay lunettes of the Murray Basin (1983) indicating a saline environment subject to wetting and drying cycles is applicable to the lunettes in the Corangamite region. Application of this theory provides evidence supporting an environment oscillating between arid and humid phases, where salinity is allowed to concentrate in near-surface water via evaportranspiration during the arid periods. The link between lunettes and wet-and-dry cycles was independently made by Sprigg (1952). He noted that those lakes and lagoons of the interdune swales in the Gambier Sub-basin that had developed lunettes were subject to annual drying.

Syngenetic speleogenesis in the Pleistocene dunes near Portland began in the mid-Pleistocene (White, 1984 & 1994). The current elevation of these cave floors are above the modern watertable, indicating that the elevation of the watertable was relatively higher in the Pleistocene (White, 1984 & 1994). The lowering to today's level is mostly due to tectonic uplift. Formation of these caves during the Pleistocene requires that periods of effective precipitation exists between the arid glacials and interglacials. Within the Naracoorte caves, Ayliffe *et al.* (1998) found speleothem growth to be episodic. According to 230 Th/ 234 U dates of the speleothems, optimum conditions for their growth were achieved during cool interstadials and stadial phases, due to the balance of adequate precipitation and moderate temperature (Ayliffe *et al.*, 1998).

It is proposed here that the pluvial phases, set within the otherwise semi-arid climate of the Pleistocene, which were responsible for the bursts of speleothem growth in the Naracoorte caves, are the same phases that relate to the cyclical wetting of volcanic lakes and the intermittent development of calcrete and speleogenesis within the aeolian dune ranges of the west. The pluvials coincided with moderate climatic conditions: either stadials or cool interstadials, and may have been restricted or intensified by movement of the STA, as dictated by ocean temperature and atmospheric circulation. These generalisations apply to the glacial stages after 130 ka, which are discussed in more detail below.



Figure 3.10. Changes in global relative sea level (after Waelbroeck *et al.*, 2002; measured against today's level and including their confidence interval as the shaded zone) and the precipitation/evaporation ratio (P/PE) of the Otway Basin since 150 ka. The aridity classifications used in the P/PE ratio plot, denoted by background shading, are those of Williams and Balling (1996); the curve was constructed from values derived from quantitative (Jones *et al.*, 1998 & 2001) and qualitative statements (authors referenced in the text). Major glacial stages mentioned in the text are marked on the sea level curve, along with the marine isotope stage (MIS) number. The trends in the P/PE ratio curve in the last 10,000 years agree well with the lake level reconstruction of Lake Keilambete (Bowler, 1981).

3.3.4 Last Interglacial Maximum (130–110 ka; MIS 5e)

As discussed in detail above, the Last Interglacial (MIS 5e) was arid or semi-arid and was preceded by an arid or semi-arid glacial phase. This is depicted by the sea level fluctuations plotted against an indicative P/PE ratio curve in Figure 3.10.

Aminostratigraphy and thermoluminescence dating of various aeolian facies of the Woakwine Range show that they are mostly representative of the transgression and high sea level stand associated with the last interglacial, at approximately 150–120 ka (Murray-Wallace *et al.*, 1999; Woakwine locality is in Figure 3.9). Further, they recognise that the youngest aeolian facies of the dune is a progradational facies, deposited as the sea level began to regress after the last interglacial. This implies that aridity and carbonate productivity continued throughout the interglacial and during its waning.

Sea level during the Last Interglacial may have been similar to today, but certain parts of land, today exposed, were inundated at the time of the Last Interglacial. At the seaward base of the Reedy Creek dune range (Figure 3.9) fossil beds of *Anadara trapezia* lie approximately 10 m above current sea level (Sprigg, 1952). However, it is generally agreed that this elevation is caused, not by a transgression 10 m beyond today's sea level, but by epeirogenic uplift of the region, as evidenced by the tilting of dune ridges (Kenley 1971; Belperio *et al.*, 1995; Murray Wallace *et al.*, 2001). A potential eustatic peak of the Last Interglacial has been established as 2 m above current sea level by using the tectonically stable Eyre Peninsula as a datum (Murray-Wallace *et al.*, 1999).

The pervasive aeolian activity of the dune building represents an arid or semi-arid climate, confirmed by the absence of speleothem growth. However, the presence of several palaeosol or calcrete horizons within the Woakwine Range, (Murray-Wallace *et al.*, 1999) shows that short, intermittent wet phases did occur. Unfortunately, there is very little lacustrine evidence from the region to support the occurrence of pluvials during the Last Interglacial, perhaps due to the time frame being outside the limits of the popular ¹⁴C dating method. However, studies of the stratigraphy, using U/Th and other forms of dating, have shown that Lake Eyre was at very high perennial levels due to a strong monsoon during the Last Interglacial (Magee *et al.*, 2004). This precipiation pattern probably indicates that the STA was positioned further south of today's location, meaning that the high pressure systems of the STA may have deprived the Otway Basin of rain year round (c.f. current climate: Section 3.2).

3.3.5 Stadials and Interstadials (110–25 ka; MIS 5, 4 & 3)

Throughout this period the sea level fluctuated with regressive phases lasting longer or outnumbering transgressive ones. The eventual result was the low sea level stand of the last glacial maximum and the gradual exposure of the continental shelf.

The rapid decrease in global temperature after the Last Interglacial allowed a more balanced P/E ratio to develop in the region and consequently speleothem growth intensified (Ayliffe *et al.*, 1998). Speleothem development characterises the entire period from 110 ka to 25 ka and has more intense periods at certain times, for example, the stadial and interstadial of MIS 5b and 5c (Figure 3.10). The intensification of pluvial periods such as this may be attributable to changes in atmospheric circulation and/or the position of the STA (Ayliffe *et al.*, 1998). Throughout the end of this period (30–24 ka), the STA was in a similar position to today and the Southern Oscillation Index was strongly positive (similar to La Niña conditions), resulting in a relatively wet climate for the SE of Australia (Harrison, 1993). Although, this time was not as wet as parts of the Holocene (Dodson, 1974; Dodson, 1975); these qualitative relationships are plotted in a timeline in Figure 3.10 (30 ka c.f. 10 ka).

The pluvial of MIS 5b and 5c may also be represented in the calcrete cap of the

Woakwine range. As mentioned previously, stratigraphic correlation and dating of calcrete layers and carbonaceous palaeosols is difficult. Despite this, it is clear that one main generation of calcrete development formed a stabilising cap on the uppermost facies of the Woakwine dune range. As this range represents the Last Interglacial (Murray Wallace *et al.*, 1999), the capping calcrete most probably represents a pluvial phase which post-dates the Last Interglacial and was complete before the arid period of the last glacial maximum at 20 ka. The cap could have formed during MIS 5 (*c.* 100–80 ka) or MIS 3 (60–35 ka).

Bowler (1981; 1983) proposed that the period from 50 ka to about 30 ka was significantly wetter than today and that many lakes in southeast Australia, which he referred to as "mega lakes", were are at high levels and low in salinity. This idea is supported by the presence of unionids (freshwater clams) in lacustrine sediments of Murray Basin lakes, one such example dated at $32,750 \pm 1,250$ BP (Bowler, 1971). Further, Bowler (1981) cites evidence of this pluvial period in the form of work from other authors showing expansion of lacustrine environments at that time. Locations mentioned include: Lakes Mungo and Tyrrell (of the Murray Basin), Lake George (of the NSW tablelands), Lake Frome (of the Great Artesian Basin) and Lake Leake of the Otway Basin.

There is evidence, including that provided above, which indicates this climatic pattern probably extended as far south as the Otway Basin at that time. Although some lakes initiated in the Pliocene in association with the eruption of great volumes of lava, some evidence exists of formation of new lakes around this time. For example, the initiation of Lake Keilambete (approx. 60 km west of Colac) occurred before 29 ka (Bowler & Hamada, 1971). Dodson (1975) found that the basin of Lake Leake (30 km NW of Mt Gambier) began to fill sometime before 50 ka and between 50 and 35 ka it was permanently wet and/or swampy, although lake levels were lower than Holocene levels. This trend is generally supported by palynology from the nearby Wyrie Swamp (Dodson, 1977). At about 35 ka the average annual precipitation for that region, based on the floral assemblage, was no more than 650 mm (Dodson, 1975). This is similar to, or slightly lower than, contemporary annual average precipitation (data from 1961–1990, Bureau of Meteorology 2005a).

3.3.6 Last Glacial Maximum (LGM; 20 ka; MIS 2)

Following the stadials and interstadials, the globe experienced continued cooling and sea level regression, culminating in the Last Glacial Maximum (LGM) at about 20 ka. This period has long been associated with a semi-arid to arid climate in parts of southeast Australia, partly due to increased continentality, caused by the regression, and partly due to decreased precipitation which resulted from the lower temperatures. Despite aridity in the southern margins, Harrison (1993) found that some lake levels in the interior of southeast Australia showed high levels between 20 ka and 12ka. This was presumably due to the monsoon having migrated further south at that time. Most palaeoclimate simulations of the LGM show a more southwards position (relative to today's position) of the STA and an associated shift of the westerlies which dominate to the south of the STA (Harrison, 1993). If these simulations are correct the westerly belt which delivers precipitation would have resided too far south of the Otway Basin to provide effective precipitation around the time of the LGM. There are several lines of evidence, outlined below, to support the idea of a semi-arid Otway Basin during the LGM. In addition, quantitative constraint of the sea level regression is important to gauge the impact on hydraulic gradients within the aquifers and is also expanded upon. The relationship between climate and sea level at this time is plotted in Figure 3.10.

There are many examples of pre-existing lacustrine environments which exhibited reduction during the LGM. In the Murray Basin, widespread dune development was underway by 18 ka and many lakes were dry by 15 ka (Bowler, 1971; Bowler, 1983). Dune development is linked to low and fluctuating lake levels and, in the case of clay lunettes, saline waters. In the Otway Basin, an indication that primary dolomite has been precipitating at Lake Beeac (10 km north of Colac) since before 10 ka (De Deckker and Last, 1989) necessitates high-salinity water, most probably produced by reduced influx to the lake. Lake Keilambete, west of Colac, began to dry out around 18 ka and was probably only an ephemeral swamp most of the time until 15 ka (Bowler and Hamada, 1971). Soil horizon development during that time indicates it was dry for several thousand years (Bowler and Hamada, 1971). Further west, at Wyrie Swamp, S.A., the driest period for the last 50,000 years occurred between 26 ka and 11 ka (Dodson, 1977). Also, lithofacies of and pollen distribution in lacustrine sediment from nearby Lake Leake, confirms that the LGM was certainly a dry period for that region (Dodson, 1975). Persistence of Euclyptus species there throughout the LGM provides a minimum annual average precipitation of 200–250 mm and minimum annual average temperature of 10° C (Dodson, 1975).

Kershaw *et al.* (2004) produced a regional synthesis for the palaeoecology of the western plains of Victoria based on palynological and sedimentary data along with ¹⁴C dates from eleven lakes in the district. This confirmed a depleted rainfall during the LGM with probable estimates being 600–700 mm annually. Despite being drier than today, these values are much higher than those deduced from the west of the basin by Dodson (1975). For the period between 20 ka and 15 ka, Kershaw *et al* (2004) infer a steppe grassland ecology with some woody shrubs and almost no trees. The low temperatures and precipitation of this time alone are not sufficient to warrant the disappearance of trees from the environment (Kershaw *et al.*, 2004). Therefore, other factors, possibly some relating to the hydrologic balance, must characterise this period. During this time,

all but three of the eleven lakes studied by Kershaw *et al.* (2004) were dry. The three wet lakes were ones that were fed by rivers or were unusually deep.

The above evidence indicates that a decline in the watertable elevation, i.e. a deepening of the unsaturated zone, had occurred at this time and was contributing to the decline of lake levels and disappearance of trees. The change in the watertable can be linked to both the decreased P/PE ratio and the regression due to glaciation (Figure 3.10). The reduced P/PE ratio would have affected infiltration and deep drainage, effectively precluding or lessening recharge to the watertable and causing a decline in pressure head. Additionally, as sea level regressed, a responding decline would have been required in the aquifers hydraulically connected to the ocean. Although this was originally a localised effect, i.e. coastal, pressure decline would have been transmitted inland with ongoing regression.

In relation to sea level change due to glaciation and oceanic cooling during the LGM, many studies concur on a major regression to a level about 120 m below that of today's sea level (Figure 3.10). Yokoyama *et al.* (2001a) found, based on microfossil evidence and ¹⁴C dating of sediments from the northwest shelf of Australia, that the low sea level of the LGM was about 125 m below current mean sea level and lasted from 22 ka to 19 ka. Subsequently, a rapid rise in sea level occurred. Waelbroeck *et al.* (2002) presented a palaeoecological, ¹⁴C and δ^{18} O study of benthic foraminifera from Atlantic and Pacific Ocean samples. From this evidence, they proposed a continuous relative sea level history for the last 450,000 years (Figure 3.7). Their work showed a rapid relative sea level rise at around 20–19 ka from -130 m to a point close to modern sea level by 6 ka. From a different methodological perspective Peltier & Fairbanks (2006) analysed corals from Barbados which further confirmed the -120 m sea level, although they preferred a maxima timing of 26 ka rather than 20 ka.

3.3.7 Latest Pleistocene (15–10 ka; between MIS 2 & 1)

In this early post-glaciation period the globe experienced rapid transgression (Figure 3.10) and warming. Glacio-hydro-isostatic numerical modelling in the northwest shelf of Australia shows that the rise in sea level after 19 ka was due to rapid deglaciation, rather than tectonism (Yokoyama *et al.*, 2001a). This period marks a transition in the Otway Basin from cool aridity and high continentality during the LGM to renewed coastal proximity and a warmer, wetter climate in the Holocene.

This general model is reinforced by observations at Lake Leake (Dodson, 1975) and Lakes Bullenmerri, Gnotuk and Keilambete (Jones *et al.*, 1998). During the transition phase the annual average rainfall around Lake Bullenmerri was approximately 400 mm, resulting in very low lake levels (Dodson, 1979; c.f. today's 800 mm). In some areas, for example, Wyrie Swamp (Dodson, 1977), the latest Pleistocene remained semi-arid throughout and increased precipitation only occurred after 10 ka.

From 15 ka to about 12 ka palynological evidence shows a reduction of woody taxa in the Otway Basin region, denoting a low P/E ratio (Kershaw *et al.*, 2004). P/PE ratios at this time ranged from 0.75 to 0.70 (Figure 3.10), despite this, the latest Pleistocene was a slightly less arid period than the preceding LGM (Jones *et al.*, 1998). Lunette formation was particularly active throughout this period, indicating some aridity and that saline environments had developed significantly over the LGM. Supporting this, Lake Keilambete's sediments and foraminifera show that a permanent, hypersaline lacustrine environment was sustained there for a significant time after 14.3 ka (Cann & DeDeckker, 1981).

Sediments aged between 12 ka and 10 ka present various forms of evidence to support an increase in both temperature and precipitation during that time. Amongst these are the increased abundance of trees (e.g. eucalypts and casuarinas) and a general increase in woody plants, changes in grass types and relatively little activity in lunette sedimentation (Kershaw *et al.*, 2004). Despite the increase in rainfall, the P/PE ratio remained low (0.7)until after 10 ka (Jones *et al.*, 1998; Figure 3.10). This evidence shows that the previously arid landscape was only now becoming wetter, despite the fact that the glacial period had begun to subside 8000 years before. Hence, there is a lag between the onset of glacial ice melting and the occurrence of water freely available in the environment. Whether the lag is directly a result of the melting or more to do with migration of weather patterns or changes in atmospheric circulation remains to be determined. Nonetheless, it seems this lag, and associated aridity, allowed for the accumulation of cyclic salts near the surface and later inundation of these areas would mobilise the salt.

3.3.8 Holocene Climatic Optimum (HCO; 9–6 ka; MIS 1)

The trend of slowly increasing rainfall continued more quickly in the Holocene until about 7 ka (Kershaw *et al.*, 2004) and, in response, the P/PE ratio increased from 0.7 to 1.1 between 10 ka and 7 ka (Jones *et al.*, 1998; Figure 3.10). After 7 ka, despite there being very little change in the vegetation patterns of the region (Kershaw *et al.*, 2004), lake levels generally shallowed a little and continued to oscillate (Jones *et al.*, 1998).

Many areas were still fairly dry at the end of the Pleistocene and the rapid humidification in the holocene had the effect in some closed lakes of greatly decreasing the salinity by the early Holocene (by 8 ka). Chivas *et al* (1985; 1986) observed this same occurrence at Lake Keilambete via the proxy of Sr and Mg ratios of fossils. They found that lake salinity remained low, and hence, levels are deduced to have been high, until around 4.2 ka. This is generally congruent with Bowler's (1981) palaeosalinity profile of Lake Keilambete gleaned from grain size analysis and those of Lakes Gnotuk and Bullenmerri (created from fossils; DeDeckker, 1982).

Dodson & Wilson, (1975) found lacustrine expansion was dominant between 7.5 ka and 5.2 ka also at Marshes Swamp, South Australia. This period was characterised by thick peat build up from a permanent, fresh and nutrient rich body of water. The underlying clay denotes the drier climate before 8.5 ka. The nearby Lake Leake and Wyrie Swamp reveal a similar pattern, with maximum lake depth, deeper than that of today, achieved around 6.9–5 ka (Dodson, 1974; 1977).

With the great body of evidence of lacustrine expansion during this time, Dodson (1974) postulated that the whole Otway Basin region had a wetter climate than today. However, later he chose to indicate precipitation levels were similar to today's (i.e. appoximately 800 mm; Dodson, 1979). The first reason for the lacustrine expansion was an increase in the P/E ratio that was driven by the warmer climate of the Holocene and greater proximity to the ocean due to sea level rise (Dodson, 1974; 1979). Second, during the HCO, the position of the STA and the associated rain-delivering westerlies to its south were probably further north of their current positions, causing the Otway Basin to fall within the year-round westerly zone (Harrison, 1993).

However, the rise of lake levels during the Holocene was probably related not only to a change in the P/PE ratio, but also to a rise in the watertable. The mechanism behind this relationship is much the same as that described in Section 3.3.6, where the decline of lakes during the last glacial maximum was partially attributed to a fall in the watertable. As sea level rise and high P/PE ratios persisted in the first half of the Holocene, both the watertable and lake levels would be expected to be high.

A simplified view of relative sea level change since 20 ka would relate continued transgression to present conditions from the time of the low stand during the LGM. However, significant evidence exists to support the idea that sea level transgressed globally to a level significantly higher than present conditions around 6 ka and subsequently regressed to today's levels by 3 ka. It has remained relatively stable since that time.

Evidence from the Otway Basin supporting this idea of a Holocene high-stand was first touched upon by Sprigg (1952), who recognised a transfersion discribed as the "Osborne high sea level phase", during which the Coorong lagoon was openly connected to the ocean, despite the existence of the dune range that is now the Younghusband peninsula, and indundation occurred up to the sea-ward base of the Woakwine Range (Sprigg, 1952; Figure 3.9).

Indicators of a mid-Holocene relative transgression of the order of 1 m to 6 m above current sea level are found in many locations around the globe, e.g. Japan (Yokoyama *et al.*, 1996), South Australia (Belperio *et al.*, 1995), northwestern Australia (Yokoyama *et al.*, 2001b), Euorope (Edwards, 2006) and the South Pacific (Moriwaki *et al.*, 2006). Timing of the stabilisation of the sea level in the region of southeast Australia is constrained to 7–6 ka (Thom & Roy, 1985; Belperio *et al.*, 1995).

Belperio *et al.* (1995) present data from five Holocene sedimentary sequences around the coast of South Australia which show apparent transgression beyond today's sea levels at around 7–6 ka and subsequent regression to modern conditions. The maximum height of the relative sea level change in those five locations ranges from 0.4 m to 4.5 m above the present low water datum. This variation is proportional to the distance from the continental slope and is therefore likely to be a response to isostatic loading on the continental shelf (Belperio *et al.*, 2002). Similarly, the global variability of the Holocene high sea level event, of both its magnitude and timing, are attributed to the variable influence of isostatic rebound, controlled by location (Lambeck, 2002).

3.3.9 Post Holocene Climatic Optimum (6 ka–present; MIS 1)

As discussed above, sea level may have been rising due to ocean thermal expansion but concurrent isostatic rebound has produced an apparently stable sea level for most of the remaining Holocene (Lambeck, 2002). Although, in the last century (1900–2000), global sea level has risen, on average, at a rate of 0.17 m/100 yrs (IPCC, 2007). The climate throughout this time remained warm and wet until about 5 or 4 ka when it experienced some minor drying and some fluctuations.

Several closed lakes in western Victoria record an increase in salinity levels between 5 ka and 2.5 ka, reflecting reduced lake levels (Bowler & Hamada 1971; Bowler, 1981; Chivas *et al*, 1985; 1986). This lowering of water levels was most likely due to a decrease in the P/PE ratio. Jones *et al.* (1998) calculated that the P/PE ratio decreased from 1.0 at 5.5 ka to 0.78 at 2.5 ka (Figure 3.10). This trend was also detected in the Gambier Sub-basin (Dodson & Wilson, 1975), to the point where complete drying and deflation of lake beds may have occurred.

After 2 ka, lake levels rose, sometimes to levels higher than today (Bowler & Hamada, 1971) and there was an associated decrease in salinity (Chivas *et al.* 1985). Jones *et al.* (1998) propose that the P/PE ratio plateaued at 0.95 until around 160 years ago. In the last 200 to 160 years, roughly the time of European settlement of Australia, lake levels have fallen and salinity levels have risen (Bowler & Hamada, 1971; Chivas *et al.*, 1985; Jones *et al.*, 1998). Jones *et al.* (2001) attribute this change to a pre-1863 climatic adjustment of the P/PE ratio (from 0.94 to 0.79) rather than European land-use changes.

3.4 Synthesis

The modern climate of the Otway Basin is Mediterranean, with rain in the cool months and hot, dry summers. The seasons are dictated by latitudinal shifts of the subtropical anticyclonic belt (STA) and the chain of low pressure systems associated with the ascending arm of the Ferrel cell. Long-term climate changes have also been caused by such shifts. For example, during the dry period of the last glacial maximum, the high pressure systems of the STA were probably positioned south of their current location, depriving the region of winter rain.

The past climate of the Otway Basin fluctuated with glacio-eustatic changes. During glacial and interglacial maxima, the climate was arid or semi-arid without significant effective precipitation. Cyclical wetting of the environment occurred between these events in the form of pluvial phases at moderate sea levels, e.g. stadials or interstadials. The most recent period like this, discounting the humid climate of the Holocene, was between approximately 110 and 30 ka. Lacustrine expansion, speleogenesis and speleothem growth all characterise this time, indicating that precipitation was effective and that, in response, the watertable was shallow. As the sea level regressed towards the low stand at -120 mAHD during the last glacial maximum (LGM), lakes contracted and the watertable probably deepened in response to a change in the P/E ratio; the sea level drop; and some tectonic uplift. The aridity of the LMG allowed salinity to accumulate at the surface via evapotranspiration. These solutes were then re-mobilised in the humid phase of the Holocene climatic optimum (HCO; 9–6 ka), during which some freshening and deepening of saline lakes occurred. Transgression to today's sea level and development of a semi-arid climate of the current interglacial had occurred by about 5 ka.

Some respite from this dry climate was observed for the period from approximately 2000 to 200 years ago, when lake levels were higher than today. About 200 years ago, a drying trend began and is still visible today. Lake levels have declined, partly in response to an increase in evaporation from climbing temperature, while the sea level has risen slowly. These patterns are forecast to continue in the future.

Chapter 4

Physical Hydrogeology

4.1 Introduction

In this chapter hydraulic evidence of groundwater flow in the Otway Basin is presented. Background information from literature and reports has been analysed and forms the basis of this introduction. First, the physical hydrogeology research issues are outlined, which link the results from this chapter with the aims of the research. Second, the evidence and conclusions of previous physical hydrogeological studies in the area are briefly summarised. Third, an overview of the known hydraulic properties of the various hydrostratigraphic units is provided.

A detailed analysis of the physical hydrogeology of the Otway Basin follows, driven by interpretation of the potentiometric maps created throughout this study. Initially, the shallow groundwater flow patterns are discussed and are assessed in close relation to the proximal surface water features. In addition, reference is made to the number of shallow aquifers as a way of describing local-scale and intermediate-scale aquifer interaction. The potentiometric maps for discrete aquifers are then presented with qualitative and quantitative examination of the groundwater flow depicted therein and comparison and contrast made to previous studies.

When discussing the flow of groundwater in the aquifer systems, a conscious effort has been made to incorporate an understanding of flow at variable scales. The relationship between flow systems of differing scales within the one groundwater catchment is well explained by Tóth (1963) (Figure 4.1). The factors that affect which scale of flow may dominate or how the differing systems may be balanced include the depth and length of the catchment, the overall slope of the topography and the amplitude of local scale topographic undulations. Importantly, such theoretical work has shown that local flow systems can have flow lines in opposing directions to those of regional scales. The former are more complex and the latter are more uniform.



Figure 4.1. Diagrammatic representation of regional, intermediate and local flow systems after Winter *et al.* (1998), modified from Tóth (1963). Flow lines represent flow in an isotropic medium (Tóth, 1963).

Further to the concept of scale variability, special attention has also been paid to the interaction between groundwater and surface water. Local flow systems have the greatest interaction with surface water features (Winter *et al.*, 1998), hence, Section 4.3 contains the majority of observations regarding surface water. However, surface water interaction with groundwater of intermediate or regional scale flow systems certainly occurs, as shown in a schematic cross section of flow in the Atlantic Coastal Plain in Georgia, USA (Figure 4.2). The Atlantic Coastal Plain is considered a basin favourable for comparison with the Otway Basin. This is due to its coastal proximity, deep extent of sediment accumulation, structural setting, age, and high quality deep groundwater. Therefore, some of the relationships between local and regional groundwater flow with surface water features identified in the Atlantic Coastal Plain may be applicable in the Otway Basin.

4.1.1 Research Issues

In the Otway Basin, groundwater occurrence and flow, dictated by climatic, topographic, structural and lithofacies controls, have been affected by anthropogenic utilisation of the resource for irrigation and urban supply from the late 1950s. Distortions in the flow regime have resulted; these may be cones of depression or loss of artesian flow due to groundwater development. In order to estimate the extent of the human "footprint", it is integral to understand the natural recharge and discharge mechanisms influencing the system. The spatial trends and processes of recharge and discharge are further constrained by taking



Figure 4.2. Diagrammatic representation of variable scale flow systems with surface water interaction in the Atlantic coastal plain, Georgia (after Priest & Clarke, 2003).

into account the nature of groundwater-surface water interaction, inter-aquifer interaction and varying precipitation and evaporation contributing to fluxes. The postulated offshore discharge from the confined aquifers, and the theories of diffuse recharge and direct infiltration are examined.

Future climate change (possibly anthropogenic) and past climate change (relating to glacio-eustatic cycles) alike result in changes to the hydrological cycle. Two particular factors affecting groundwater resources in the Otway Basin are addressed in this study. The first is, the changes to recharge patterns as a result of climate change and the second is the sea level fluctuations and how they influence the hydraulic balance and hydraulic gradients onshore.

4.1.2 Previous Work

This chapter concentrates on groundwater flow in the more hydraulically conductive hydrostratigraphic units, namely, the Plio-Pleistocene Heterogeneous, Pliocene-Quaternary Volcanic, Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers and, where possible, the Upper Cretaceous Aquifer. The Upper Tertiary Aquitard is discussed in a limited context. The four uppermost aquifers contain high quality groundwater that is widely used for irrigation, domestic and urban supply, stock and dairy use and industrial use. Although parts of the Upper Cretaceous Aquifer also contain a high quality water resource, it is not exploited as extensively because it is less accessible in most areas. Being such an important resource, the groundwater of these aquifers has been the focus of various other studies.

Several authors have produced watertable maps and potentiometric surface maps for a number of the aquifers of the basin, usually within specific sub-basins, and sometimes limited to one hydrostratigraphic unit. These studies developed the understanding of the flow paths of groundwater in the basin and identified potential recharge and discharge zones (Ward, 1941; O'Driscoll, 1960; Kenley, 1962; Colville & Holmes, 1972; Waterhouse, 1977; Blake, 1980; Williamson & Turner, 1980; Leonard, 1983; Leonard et al., 1983; White, 1984; Nahm, 1985; Love et al., 1992b; Love et al., 1993; Grimes, 1994; Witebsky et al., 1995; Coram, 1996; Adler, 2003; Bennetts, 2005; Petrides & Cartwright, 2006; Raiber, 2008). In addition, the Border Zone reports regularly produce hydrogeological maps spanning the region on either side of the state division and, most recently, a series of hydrogeological maps, commissioned by Southern Rural Water, were completed by SKM in collaboration with data from this study (SKM, 2009a). Generally, groundwater flow in the confined aquifers of the Otway Basin is from the northern margins, where regional recharge occurs, towards the ocean. Flow in the unconfined aquifers mimics this path, except in the case of localised flow paths which interact with surface water features and can be quite variable.

Some early descriptions of recharge to the aquifers at the basin margins involved analysis of hydrographs and a link was developed between recharge and the precipitation peak in winter, along with a lag in the the hydrograph response (e.g. O'Driscoll, 1960). Discharge was documented from early times to be near the coast and/or into the ocean, including descriptions of flow paths of the groundwater and offshore or coastal discharge springs, sometimes with volumetric flow rates (Tate & Dennant, 1895; Sprigg, 1952; Wylie, 1969; Waterhouse, 1977; Shugg, 1984).

Many studies present data in the form of aquifer characteristics from pump tests on particular wells or porosity estimations and this information is used in this study when assessing groundwater flow (Harris, 1969; Morton, 1969; Thompson, 1972; Blake, 1974; Waterhouse, 1977; Shugg, 1976; Blake, 1980; Shugg, 1981; Leonard, 1983; Gill, 1989; Love *et al.*, 1991; Shugg, 1993; Bennetts, 2005).

Quantitative studies of groundwater flow in regions of the basin followed the establishment of aquifer characteristics and involved both analytical solutions and flow models. Holmes & Colville (1970a and 1970b) made estimates of recharge in areas of varying land use by employing physical hydrogeology methods. Numerical models of physical groundwater flow, which in most cases relied upon previously reported hydraulic characteristics of aquifers, were created by Harrington *et al.* (1999); Duran (1986); Skidmore (1992); Walker *et al.* (1992); Dogoli (1993); Stadter & Yan (2000); SKM (2005); Smitt *et al.* (2005) and Cox *et al.* (2007).

4.1.3 Overview of Aquifer Properties

The physical characteristics of the hydrostratigraphic units have been obtained via research and correlation of unpublished reports and published data, including pump tests, porosity and grain size data. The main difficulty with interpreting information from individual pumping tests is to extrapolate these values to representative values at the regional scale.

The major publications and reports with values of hydraulic parameters of the aquifers and aquitards of the basin were reviewed and the data were collated to create a summary (Table 4.1). This is not a definitive collection of data for the tabulated parameters; the region is of such a size that significant amounts of knowledge have been accumulated and could not be condensed into one page. Some figures in Table 4.1 are representative averages reported by the authors and others relate to single formation tests in one well. This difference is important, especially when considering the variability of hydraulic properties of the Upper Tertiary Carbonate Aquifer caused by karstification (O'Driscoll, 1960; Blake, 1974).

The provenance of the information requires consideration when using the data in regional scale interpretations. Schaeffer (2008) concluded from a detailed review of the applicability of the results of individual pump tests to an aquifer at a regional scale that upscaling local hydraulic data was only appropriate or representative if it incorporated a range of methods of determining hydraulic characteristics, e.g. grain size analysis in addition to pump tests. Further, characteristics that were tested on local-scale calibrated groundwater flow models are more reliable. Calibrated models in the Otway Basin include those of Harrington (1999) in the Gambier Basin; Skidmore (1992) in the Port Campbell Basin and Witebsky *et al.* (1995) and Teng (1996) in the Barwon Downs Sub-basin.

Specific yield or storage coefficient (S) estimates are largely absent for the Pliocene-Quaternary Volcanic Aquifer because this parameter is difficult to determine (Table 4.1). This is due to two reasons. First, that porosity is very variable due to the secondary porosity being most important in terms of flow; and second, that recharge areas behave as unconfined aquifers and flow-thru areas behave as confined aquifers (Thompson, 1972). The latter reason is attributed to the confining effect of the clay layers that represent palaeosol horizons between lava flows. Recharge zones are often eruption points where layering of lava flows has not developed.

In addition to the information in Table 4.1, ranges of porosity values that represent the spread of data for particular formations in offshore wells is presented (Table 4.3). The data ranges were obtained from PORPERM, a database administered by Geoscience Australia which houses data relating to oil and gas reservoirs and traps of major Australian sedimentary basins.

Sources are acknowledged by superscript numbers: ^[1] Harrington *et al.*, 1999; ^[2] Love *et al.*, 1993; ^[3] Love *et al.*, 1994; ^[4] Blake, 1974; ^[5] Shugg, 1981; ^[6] Shugg, 1981; ^[6] Shugg, 1981; ^[6] Shugg, 1982; ^[14] Morton, 1972; ^[12] Harris, 1969; ^[13] Shugg, 1993; ^[14] Morton, 1969; ^[14] Love *et al.*, 1994; ^[4] Blake, 1994; ^[4] Blake, 1980; ^[14] Morton, 1969; ^[14] Love *et al.*, 1994; ^[14] Coram, 1996; ^[14] Shugg, 1981; ^[14] Morton, 1969; ^[19] Success are acknowledged by superscript numbers: ^[14] Love *et al.*, 1994; ^[14] Shugg, 1983; ^[14] Morton, 1969; ^[14] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Morton, 1969; ^[15] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Morton, 1969; ^[15] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Morton, 1969; ^[15] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Morton, 1969; ^[15] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Morton, 1969; ^[15] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[15] Love *et al.*, 1984; ^[14] Shugg, 1984; ^[14] Morton, 1986; ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Morton, 1969; ^[15] Coram, 1984; ^[14] Shugg, ^[14] Shugg, 1984; ^[14] Morton, 1969; ^[14] Shugg, ^[15] Coram, 1986; ^[14] Shugg, ^[14] Shugg, ^[14] Shugg, ^[14] Shugg, ^[16] Coram, 1986; ^[14] Shugg, ^[14] Shugg, ^[14] Shugg, ^[14] Shugg, ^[16] Coram, ^[15] Shugg, ^[14] Shugg, ^[14] Shugg, ^[16] Shugg, ^[14] S hydraulic conductivity (standardised to 20°C, horizontal unless stated as vertical in superscript); Sub-basins: G—Gambier, T—Tyrendarra, PC—Port Campbell. Table 4.1. Hydraulic characteristics for the different aquifers and aquitards of the Otway Basin. n_e —effective porosity; S—storage coefficient in confined aquifers, specific yield in unconfined aquifers; Q/s-specific capacity (discharge per unit drawdown); Trans.--Transmissivity; b--aquifer thickness or thickness tested; K-citation.

Aquiter	Lithology/lithofacies	Sub-basin: Region <i>[borehole]</i>	n _e [%]	S [dimensionless]	Q/s [m ³ /day/m]	Trans. [m³/day/m]	م [۳]	K [m/day]	Typical yield [L/sec]
	ash and tuff (Tower Hill) basalt	T: Warrnambool region	5 [4] 5 [4]			$\frac{450}{180}^{[4a]}$	48 ^[4] 61 ^[4]	$rac{9.45}{2.93}^{[4a]}$	$1.26-12.6^{[4]}$ 0.63.6.3 ^[4]
PQV Aquifer	scoria all lithofacies combined basalt	T: Condah Region	$5^{[18]}$				5	0.4 ^[18] 6.0 ^[17]	
	scoria	1 - - 5 5					$9^{[16]}_{116]}$	48 ^[16]	
	later phase basalts	PC: Corangamite Region					6 [16]	10 ^[16]	
	earther phase basalts all lithofacies combined		[1] 2 <	0.05 $^{[7]}$		$450^{[7]}$	13 -	4 100	
	quaternary clays	PC: Corangamite region					8.3 ^[16]	$3.1^{[16]}$	
	quaternary sands	PC: Corangamite region	3				$7.3^{[16]}$	$34^{[16]}$	3
	Bridgewater Formation	T: Warrnambool region	$10^{[4]}$	[19]		23 [4a]	48 [4]	$0.5^{[4a]}$	<1.26 ^[4]
PPH Aquifer	Bridgewater Formation and others	G: Robe region T. Condob Domion	10 [18]	0.09 - 0.18		2007 - 2713 [12]		A 2 [18]	
	Moorabool Viaduct Sand	T: Warrnambool region	$10^{[4]}$			$36^{[4a]}$	${}^{30}_{9}{}^{[4]}_{16]}$	$\frac{4.5}{1.2}$ [4a] 4.5 [16]	<1.26 ^[4]
		G: Mt Gambier region	2	0.1 - 0.4 ^[10]			5	7 - 60 [10]	
	Gambier Limestone	G: general G: Northern region	50-60 [9] 30 [9]	0.2 [9]		200 - 10000 ^[2]			
		G: Robe region	2	1		$255 - 1798^{[12]}$			
UTC Aquifer	Gambier Limestone (cavernous)	G: Mt Gambier region	[0] 00 00	[1]	[9]	[1]		$\frac{110}{2} \cdot \frac{270}{61}$	
	Port Cambbell Limestone	PC: Warrnambool region	$20-30$ [13] $40^{[13]}$	$\begin{array}{c} 0.1 & - 0.2 \\ 0.2 & \left[13 ight] \end{array}$	417 ^[2] 20 - 200 ^[13]	186^{13} 1000 - 9000 ^[13]	$300^{[13]}$	$\frac{7.2}{7}$ ^[13]	
		T: Condah & Hawkesdale	5 ^[11]				3	0.1 - 1.5 [18 & 19]	3
	Port Campbell Limestone (cavernous)	T: Warrnambool region	20			60 ^[48]	$183^{[4]}$	$0.3^{[4a]}$	<25 ^[4]
UT Aquitard	Gellibrand Marl	PC: Corangamite region					$42^{[16]}$	9.5 ^[16]	
UTC Aquifer	Clifton Formation	T: Condah Region	$18^{[18]}$					3 - 18 ^[18]	
UT Aquitard	Narrawaturk Marl	G: Southern region						10 ⁻⁷ - 0.001 ^[2 vert.]	
		G Northern region	15 ^[1] 15 of ^[3]						[8]
		G. Mt Gambiar ramon	07-0T	0 0001 [10]		180 - 1600 ^[10]		U.Y - J.Y, AVE 2.JY	
		G. MU GAIIDIEI LEGIOII G. general	20.30 ^[2]	TOOOO		200 - 1600 ^[2]	100-400 ^{[2}	a] 28 ^[2a]	
		G: [Naracoorte 8]				1244 $[14]$	$30^{[14]}$	$\frac{1}{42}$ ^[14a]	
		T: Portland region		$0.0001^{[6]}$		$700 - 1000^{[6]}$	6	$13^{[6]}_{\mathrm{[ea]}}$; 17 - 24 $^{[6a]}$	
		T: [Portland 13]		0.000094 ^[9]		1383 ^[o]	37 ^[0]	$20^{[60]}$	
	Dilywn Formation, Eastern View	T: [Heywood 14] T: [Portland 11]		0.00495		376 ^[0] 546 ^[20]	16 [0]	17 ^[04] 19 ^[20]	
LTS Aquifer	Formation and sandy lithofacies	T: [1 on turner 11] T: [Belfast 13]		0.000547 ^[8]		$^{040}_{1216}$	$55^{[8]}$	$\frac{10}{14}$ [8a]	
	of the Mepunga Formation	T: [Homerton 4]				$70^{[20]}$		$6.9^{[20]}$	
		T: [Mouzie 1] T: Warrnamhool region	1 E [4]			61	769 [4]	2.1^{-2} 0 13 [4a]	<196 [4]
		PC: [Paaratte 2]	0	$0.000063^{[8]}$		$411^{[8]}$	$26^{[8]}$	11 ^[8a]	
		PC: Warrnambool region		[8]		[8]	[8]	$5 - 10^{10}$	
		BD: Geelong Kegion BD: <i>(Gerangamete 16</i>)		0.0003		306 - 512 ^[8]	$\frac{40}{75}$ ^[8]	9 5 ^[8a]	
		BD: [Murroon 23]		0.000016 [8]		64 [8]	13 [8]	2 [8a]	

Hydrostratigraphic Unit	Porosity Range
Lower Tertiary Sandy Aquifer (Dilwyn Formation)	0.25-0.4
Lower Tertiary Aquitard (Pember Mudstone)	< 0.10
Upper Cretaceous Aquifer (Pebble Point Formation)	0.25 - 0.4
Upper Cretaceous Aquifer (Timboon Sand, Paaratte Formation)	0.2 - 0.3
Upper Cretaceous Aquifer (Flaxman Formation, Waarre Formation)	0.05 - 0.25

 Table 4.2.
 Porosity ranges of aquifers and aquitards of the Otway Basin from PORPERM database,

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4.2 Methods

This chapter presents original work in the form of watertable or potentiometric surface contour maps for the Otway Basin aquifers. The 3D computer model created in GoCAD, as distinct from a numerical model, was used to integrate a variety of data sets to create these maps. The program was used to perform inverse distance weighting interpolation (IDW) on the data points to create the potentiometric surfaces. Below is a detailed methodology which identifies the data used and describes the process of readying the data for the applications.

4.2.1 Data Sources

The data presented in this chapter are mainly measurements of groundwater levels in bores screened in various aquifers and aquitards from the South Australian and Victorian sections of the basin. However, the supporting data relating to bore construction, location and the date of measurement are also important.

Sinclair Knight Merz (SKM) is responsible for maintaining the Victorian groundwater database (GMS) and responding to data requests on behalf of the Victorian Government. Raw hydrogeological data were collected from SKM for the Victorian section, including depth of screened interval and hydraulic head data. The data sourced from this database had coordinates recorded in eastings and northings (grid data) in the AMG66 datum (geocentric). These data were converted to the Map Grid of Australia 1994 (MGA94) using *GDAit*, a program developed by the University of Melbourne (http://www.sli.unimelb.edu.au/gda94/).

Primary Industries and Resources South Australia (PIRSA) maintains the State Observation Well Database (OBSWELL), from which groundwater data were collected for the South Australian section of the basin. The location information for the South Australian bores was provided in both grid and geographic coordinates, in the GDA94 datum. The two datasets have been integrated with one another to produce the results of this chapter.

4.2.2 Spatial aspects of the data

The South Australian database accessed contains an aquifer name field, making the task of separating data into aquifers simple. The Victorian database does not include this information, hence it had to be derived from the fields that were provided (e.g. aquifer lithology and screen depth), in combination with the aid of the previously constructed 3D hydrostratigraphic model (Section 2.2.2).

The location information was the first data field to be examined. Many wells in the database have inaccurate or incomplete location information. These gaps needed to be filled, or otherwise the data discarded. Some wells with erroneous coordinates were discovered due to the easting and northing not falling within the area of the parish name assigned to the well. Each parish in the basin was checked and these data were discarded. The precision of many well locations was improved by an audit conducted by the state in 2006. Where available, the coordinates from the audit were used.

The poorest aspect of the location data was the elevation information. Often, the elevation field was not recorded in the database. In this case, after all location data had been converted to MGA94, the coordinates were projected onto a digital elevation model (DEM) using *ArcGIS* and the corresponding elevation was extracted for each well. The DEM used is from the SRTM (Shuttle Radar Topography Mission), which was developed by NASA (http://www2.jpl.nasa.gov/srtm/) and is distributed by the USGS (http://seamless.usgs.gov/). The SRTM DEM has a horizontal spatial resolution of 90 m in Australia. In the case where elevation was provided with the original data, it was validated using the SRTM data. Some wells have elevation data recorded in the geological database used to create the hydrostratigraphic framework. These data were also used in place of missing elevations, or as a comparison.

Once satisfactory location data were collated, the depth of the screen interval of each well could be assessed and converted to screen elevation in mAHD by subtracting depths from ground elevations. The screen information came from several data fields in the GMS which had to be amalgamated. First priority was given to the screen top and bottom depths. Many wells lacked these data, but did have depths recorded for the top and bottom of the interval where water was encountered during drilling. These fields were of second priority for providing the screen location. In the absence of these fields, it was possible to use recorded sample depths as screen interval depths. In the absence of all the preceding fields, the screen interval was assumed to be close to or at the bottom of the well and so total depth was used as the bottom of the screen interval. These wells bear a quality control flag in the table in Appendix C. Once the screen interval positions had been established, a combination of data sources were used to discover the geology at the screen depths. Some boreholes have stratigraphic or geologic logs, for example the logs used in creating the hydrostratigraphic model, hence screen geology was simply determined in these cases. Stratigraphic records were also used in the case of neighbouring boreholes, if proximity allowed. The data of second priority for determining the geology of the screen interval were the lithological information housed in the GMS. This often required interpretation and cross checking with the model. If both these data were not available for a borehole, the hydrostraigraphic model was the sole means of assigning a hydrostratigraphic unit to its particular screen depth.

The relationship between the depth to the top and bottom of the screen interval and the depth to the top and bottom of the aquifer in question was not always simple. That is to say, some screens cover a depth interval intersecting more than one hydrostratigraphic unit. Alternatively, due to the assumptions listed above, some screen intervals themselves were not well constrained, meaning that interpretation was necessary. More information about these cases can be found in Appendix C. It is sufficient to say here that single-aquifer boreholes were used as reliable data points, while overlapping wells were attributed to the main aquifer and their hydraulic data used with caution.

4.2.3 Temporal aspects of the data

Potentiometric maps for each of the regional aquifers should represent a particular snapshot in time, but due to the lack of necessary data, this is not strictly the case here. Data used to create the maps are the most recent observations for each well. In some cases, measurements were taken from investigation bores which were never completed as permanent observation bores. This means that the most recent measurement is the only measurement, which may be as old as the 1950s for some wells of the deeper aquifers. In the case of old measurements, data points that appeared anomalously low or high when compared to neighbouring points were removed and, overall, the mixture of measurement dates does not detract from the usefulness of the maps of the deep aquifers.

On the other hand, the shallow aquifers of the Otway Basin are more subject to variability and are represented by a better data set. Of the measurements used for the shallow aquifer maps, 85% were taken in the last 30 years. Wells that exhibit extreme fluctuations, like drawdown, or strong inter-annual trends have been avoided in the compilation of the maps, both for the shallow and deep aquifers.

In addition to inter-annual variation, seasonal variation is visible in hydrographs for the shallow aquifers. It is desirable to avoid misrepresenting the hydraulic conditions in these aquifers by combining measurements from different seasons. However, due to differing lag times for hydrograph response to infiltration, peaks of different wells are not entirely in phase with each other. This means that comparing measurements from different seasons is not altogether misleading. Nonetheless, the bias in the data for the shallow aquifers is intentionally towards measurements taken in summer or autumn, constituting 70% of the total.

The initial intention to produce two maps for each aquifer, one representing the predevelopment conditions (e.g. c. 1950) and another representing the current conditions, was not possible due to the paucity of data.

4.2.4 Density correction of hydraulic head

The hydraulic head data, once separated into aquifer units as described above, were used to create the watertable and potentiometric surface maps. These maps depict contours where the total hydraulic head is equal (isopotentials). Total hydraulic head (or, simply, hydraulic head) is defined by the equation:

$$h = z + \psi \tag{4.1}$$

where h is hydraulic head, z is elevation head and ψ is pressure head. Elevation head relates to the position of the aquifer being measured in relation to a datum. Pressure head relates to the hydraulic pressure at the point of measurement and is expressed as:

$$\psi = \frac{P}{\rho \cdot g} \tag{4.2}$$

where P is the groundwater pressure; ρ is the groundwater density and g is gravitational acceleration. This pressure causes the groundwater to rise to a certain height in a well, this level being the measurable component of hydraulic head and the water column in the well representing ψ .

However, as can be seen from Equation 4.2, pressure head (ψ) can vary according to water density without a change in groundwater pressure. Therefore, before water level measurements can be used to calculate h, the density of the water must be used to correct ψ so that all measurements relate to water of a standard density.

The accepted equation for density correction of hydraulic head is (Post *et al.*, 2007):

$$h_c = \frac{\rho_u}{\rho_f} (z_u + \psi_u) - \frac{\rho_u - \rho_f}{\rho_f} z_u$$
(4.3)

where h_c is the corrected hydraulic head; ρ_u is the uncorrected density; ρ_f is standard freshwater density, 1g/cm³, the benchmark against which all measurements will be corrected; z_u is the uncorrected elevation head; and ψ_u is the uncorrected pressure head (n.b. this notation is different from that of Post *et al.* (2007) but it expresses the same equation).

Density corrections for hydraulic head are often overlooked by many studies (Post *et al.*, 2007). However, this is not a trivial matter, considering that the density of a body of water depends on three quite variable characteristics: temperature, salinity and pressure (depth). The above density correction was undertaken on all hydraulic head data used in this study due to the varied salinity within certain aquifers (e.g. the Pliocene-Quaternary Volcanic Aquifer) and the depth and related temperature variations in others (e.g. the Lower Tertiary Sandy Aquifer). The density ρ_u used in the correction (Equation 4.3) was calculated using an algorithm from Fofonoff and Millard (1983) which was derived from the 1980 International Equation of State for Seawater (EOS80; UNESCO, 1981b). Use of the EOS80 required temperature (T), salinity (S) and pressure (P) values to be assigned to each well (explained in more detail below).

The EOS80 was developed using the 1978 Practical Salinity Scale (PSS; UNESCO, 1981a) and the International Practical Temperature Scale of 1968 (IPTS-68; Millard & Yang, 1993). The EOS80 was developed from experimental data for a constant pressure (Millero & Poisson, 1981) and for variable ocean depths (Millero *et al.*, 1980). The tested parameter limitations of the EOS80 are 0-42 for salinity, $-2^{\circ}C-40^{\circ}C$ for temperature and 0-10,000 decibars for pressure (Fofonoff & Millard, 1983). The EOS80 is applied here beyond these temperature limits (up to appoximately 70° C) due to lack of an alternative. Millero (2000) acknowledges that changes in the geochemical composition of the seawater and their effect on the relationship between electrical conductivity and salinity are a limitation of the EOS80. However, these changes may be considered acceptable and, given the diversity of geochemistry within the Otway Basin groundwaters, unavoidable. Millero (2000) noted the applicability of the EOS80 to estuary waters, despite its development for seawater, and this forms the platform for the application of the equation to groundwater in this study.

Salinity values (S, unitless) for the water density calculation were derived from the electrical conductivity (EC) values. Groundwater EC for each data point, where not previously known, was extracted using the interpolated electrical conductivity maps (Chapter 6). These conductivities (mS/cm) were converted to conductivity ratios (R, unitless) using the equation:

$$R = C_{(S,T,P)} / C_{(35,15,0)} \tag{4.4}$$

where $C_{(S,T,P)}$ is the groundwater conductivity and $C_{(35,15,0)}$ is the conductivity of standard seawater with a salinity of 35, at 15°C and atmospheric pressure. The value for $C_{(35,15,0)}$ was taken as 42.9140 mS/cm (Culkin & Smith 1980). Salinity values were then calculated from R values, using the practical salinity scale (PSS78)(Fofonoff and Millard, 1983). The PSS78 is a defined ratio between the measured conductivity of the sample and a standard solution of KCl at constant temperature and pressure (Fofonoff, 1985). By the PSS78 definition, a sample has a salinity of 35 if it has the same conductivity as a KCl solution of concentration 32.4356 g/kg. Hence, there are no units for PSS78 salinity, however, many users describe these salinity values as having the unit PSU (practical salinity unit) to avoid confusion with other comparable, although different, measures of salinity (e.g. parts per thousand; ppt).

Temperature values (T) for the water density calculation were derived for each bore by using the depth corresponding to the bottom of the well's screen interval and applying a regional average geothermal gradient of 3.6°C per 100m (King, 1987). Various methodologies exist for determining geothermal gradients from heat flow data (Cull, 1982). In the Otway Basin, maps of variable geothermal gradient have been developed from several methods, including use of formation-water temperatures and corrected bottom hole temperatures (BHTs)(King, 1987; SKM, 2005b; Driscoll, 2006). The region has a range of vertical geothermal gradients from 3°C to 7°C per 100m, with an average of 3.6°C, which is generally high compared to the "normal" average of 2° C to 3° C (King, 1987; Bowen, 1979). The areas of high geothermal gradient in the Otway Basin are generally characterised by shallow depth to basement and/or faults and recent volcanism. Overburden lithology is also has been found to correlate with areas of elevated formation water temperature, e.g. in the Gippsland Basin, at depths as shallow as 520 m, an overburden comprising thick sequences of coal produced a water temperature of 70°C (Nahm, 2002). The temperature measurements taken in the field correlate well with the gradients of King (1987) and point data from other sources (e.g. Shugg, 1976), and thus validates the method for modern hydraulic head correction.

Generally, the use of the EOS80 with modern temperature data requires conversion of temperatures from the International Temperature Scale of 1990 (ITS-90) to IPTS-68 (Millard & Yang, 1993). In this study, most of the temperature values used are estimates from geothermal gradients, hence, the minor adjustment from ITS-90 to IPTS-68 is not necessary.

Pressure values needed for the water density calculation were assigned to the data points by applying a pressure gradient to the depth of the bottom of each screen interval. The gradient reflects the amount of pressure the overburden exerts on the aquifer at a certain depth according to bulk overburden density $(2.67g/cm^3)$ and acceleration due to gravity $(9.8m/s^2)$. The resulting pressure gradient is 26166 Pa/m. Overburden density is actually variable, but an heterogeneous pressure gradient could not be used because of complexity constraints. The overburden density was selected as representative because it is the bulk density of granodiorite and is characteristic of the upper continental crust (Taylor & McLennan, 1985; Rudnick & Gao, 2003).

4.3 Shallow groundwater flow

The shallow groundwater flow patterns of the Otway Basin are depicted in the watertable map (Figure 4.4). This analysis of the watertable highlights the interconnectivity of the groundwater within different aquifers. An assessment of the flow patterns of the shallow groundwater of the basin is concerned with groundwater from all the hydrostratigraphic units of the basin because each unit crops out at the surface to some extent (Figure 2.1). The flow patterns derived from the watertable are explained in the context of the local geomorphology and surficial lithologies.

In investigation of the watertable implicitly includes groundwater-surface water interaction. Areas where shallow groundwater flow is topographically driven inherently display flow patterns that mimic those of surface water flow. In addition, surface water features of the Otway Basin may be related to local groundwater flow path ends or beginnings (Figure 4.5).

Surface water features are commonly categorised by hydrogeologists according to their relationship to groundwater flow; categories are usually distinguished into groups involving recharge features, through-flow features, and discharge features. Recharge features are those that provide for water migration to the watertable via deep drainage, e.g. a losing stream. Discharge features, e.g. gaining streams, are those that receive groundwater discharge, i.e. baseflow. To enable groundwater surface water interaction, a hydraulic gradient must exist between the two bodies and the gradient's direction will determine whether discharge or recharge shall occur. These gradients are often discernible from the curve of isopotential contours close to a surface water body. For example, the isopotentials of a watertable near a gaining portion of a stream converge upstream, whereas those near a losing portion converge downstream (Figure 4.3). However, the watertable map of this study was constructed from bore data and insufficient density of information exists to create a map like that of Figure 4.3. Where paucity of data existed the watertable elevation (Figure 4.4) was adjusted to represent a subdued version of the digital terrain model.

The shallow groundwater of the Gambier, Tyrendarra and Port Campbell Sub-basins generally flows coastwards (Figure 4.4). The flow within the Barwon Downs Sub-basin is ultimately coastward, but is diverted by the Otway Ranges towards the coast to the northeast, near Geelong. Also in the east of the basin, the Corangamite region, named for Lake Corangamite at its centre (Figure 4.5) contains internally-draining flow lines. Lastly, in a region north of the Tyrendarra Sub-basin, flow is dominantly westwards (Upper



Figure 4.3. Diagrammatic representation of gaining and losing streams after Heath, (1983). Isopotential contours of the watertable are solid lines, groundwater flow lines are dashed.

Wannon region). Muddy Creek and Grange Burn are the drainage features within this region; they are both tributaries for the Wannon River, which flows along the northern boundary (Figure 4.5). The watertable dynamics of each region are discussed in turn below.

4.3.1 Gambier Sub-basin

In the north of the Gambier Sub-basin the natural surface drainage features flow southwestwards over the Kanawinka escarpment and continue through the Naracoorte dune range, culminating in the group of lagoons before the Stewart Range (e.g. Bool Lagoon, Figure 4.5; for dune locations, refer to Figure 3.9). Southwest of the range, artificial drains to the ocean have been developed both parallel and orthogonal to swales. Many swales contain swamps or permanent lakes (e.g. Lake Bonney). In the southeast of the Sub-basin the Glenelg River meanders along a deeply incised valley, reaching the coast at Nelson. Significant springs emerge from the Upper Tertiary Carbonate Aquifer (Gambier Limestone) near the coast at Ewens Ponds and Piccaninnie Ponds and along the Glenelg River below Dartmoor (Figure 4.5).

Apart from the limited volcanic rock outcrop, the surface geology of the Gambier Sub-









basin consists of the Plio-Pleistocene Heterogeneous Aquifer, Upper Tertiary Carbonate Aquifer and Lower Tertiary Sandy Aquifer (Figure 2.1). The region is overwhelmingly dominated by the two sub-aquifers of the Plio-Pleistocene Heterogeneous Aquifer, alternating in elongate zones subparallel to the coast. The Lower Tertiary Sandy Aquifer distribution at the surface is minimal. Within the Sub-basin it is constrained to the incised valley of the Glenelg River, close to the Otway Basin's northern margin. There are also small exposures on the Dartmoor Ridge where it is crossed by the valleys of the Crawford and Stokes Rivers. The Upper Tertiary Carbonate Aquifer (Gambier Limestone) is exposed in abundance in the southeast corner of the Sub-basin and in a less significant, yet still regional-scale outcrop in the upthrown block north of the Kanawinka Fault.

The Gambier Sub-basin hosts shallow groundwater flow which originates in the inland areas, towards the center of the Sub-basin and close to the Dundas Plateau, and which then flows radially towards the coast. It is difficult to definitively say where the divide exists between the Gambier Sub-basin and the Murray Basin towards its north in terms of shallow groundwater. This is because the groundwater divide is weak and consists of a shift from westwards flow in the south to northwestwards flow further north. However, it is reinforced to a certain extent by the physical division of the hosting aquifers, achieved by the shallow depth to the basement rocks at the Padthaway Ridge. The hydraulic divide bounding the eastern edge of the Gambier Sub-basin is much more pronounced. It is a groundwater ridge that coincides geographically with the Dartmoor Ridge.

The Dartmoor Ridge is a basement high, however, its location is also marked by a topographic high (Figures 1.2 and 2.2). The topographic ridge is due to two factors: normal movement along the Kanawinka Fault (refer to Figure 2.21) causing uplift in the east and incision of the Glenelg River in the west. The ridge is an important zone as the elevated land results in a small orographically-induced rainfall high over the area (Figure 3.4) that perpetuates the groundwater mound and local downward gradients. Conversely, a groundwater "gully" has developed in association with the valley of the Glenelg River. As such, these features of the shallow groundwater flow in the Gambier Sub-basin are strongly controlled by topographic features.

In contrast, some features of the watertable are the result of the underlying geology. For example, two suites of volcanoes, the maars of the Mt Gambier suite (between the 10 and 20 mAHD contours) and the scoriaceous volcanoes of the Mt Burr suite (around the 60 mAHD contour) lie in the southeast of the Gambier Sub-basin (Figure 4.4; for a map showing the types and distribution of the eruption centres, refer to Figure 7.8). The two suites are separated by a zone of steep hydraulic gradient which has been mapped and described previously (e.g. Waterhouse, 1977; Love *et al.*, 1993). However, within the geographical area encompassing each group, the hydraulic head is almost unchanging. This implies good

hydraulic connection within the suites but not between them. The high gradient zone is possibly created, as suggested by Waterhouse (1977), by changes in the transmissivity - via either changes in thickness or hydraulic conductivity - of the underlying Upper Tertiary Carbonate Aquifer, to which the surficial aquifers are hydraulically connected. Alternatively, it may have developed as a linking zone between the two proximal areas of differing pressure (O'Driscoll, 1960); the lower pressure zone perpetuated by discharge from the Upper Tertiary Carbonate Aquifer induced by faulting (i.e. Tartwaup Fault; Figure 2.21). From the formation of a groundwater mound at the Mt Burr suite, it is apparent its scoriaceous porosity and associated faults act as good conduits for recharge.

In Figure 4.4, the shallow groundwater flow lines can be seen to terminate at or near the coast, indicating coastal or near-coast submarine discharge. Discharge also occurs via mechanisms of abstraction and evapotranspiration. Evapotranspiration occurs mostly from the swales and other areas where the watertable is close to the surface, affecting mainly the groundwater that flows in the shallowest, localised flow paths. Coastal springs emanating from the shallow aquifers, e.g. the Gambier Limestone of the Upper Tertiary Carbonate Aquifer (Sprigg, 1952; Waterhouse, 1977), occur in several places in the Gambier Sub-basin (e.g. Ewens Ponds, Figure 4.5). In addition, near-coast diffuse seepage is an important contributor to submarine shallow groundwater discharge; albeit a difficult factor to quantify. The fact that the discharge from the shallow rocks of the area is at or near the coast is controlled by the sea level and the aquifers' hydraulic connection to the ocean. The interface between the fresh groundwater and the dense marine water within the aquifers slopes inland and forces groundwater flow upwards, aiding discharge. As evidence of this oceanic connection, Figure 4.4 shows that the elevation of the watertable close to the coast is close to 0 mAHD. The low hydraulic gradient near the coast, as shown by the widely spaced isopotentials, is a poor defence against inland encroachment of the freshwater-saltwater interface. Put another way, the Gambier Sub-basin is possibly the most susceptible to future rise in sea level.

4.3.2 Upper Wannon region

As described above, the surface hydrology of the Upper Wannon region consists of a generally westwards regime including the features Muddy Creek, Grange Burn and the Wannon River (Figure 4.5).

Almost the entire surface area of the Upper Wannon region is covered by the Pliocene-Quaternary Volcanic Aquifer. The exception is the westernmost corner of the area where the Plio-Pleistocene Heterogeneous Aquifer, the Upper Tertiary Carbonate Aquifer (Port Campbell Limestone) and the Lower Tertiary Sandy Aquifer are exposed. In the west, groundwater and surface water flow converges in a catchment-like pattern, enabling interaction of groundwater from the different aquifers (Section 5.3). The northern and western boundaries of the region are marked by basement outcrop.

Groundwater flow in the Upper Wannon region is mostly westwards but also involves some northwards flow from the southern boundary of the region. The southern boundary is a groundwater flow divide with the Tyrendarra Sub-basin and is marked by two major eruption centres, Mt Rouse and Mt Napier. These are composite volcanoes, having erupted both lava and scoria (Rosengren, 1994). The watertable high of the divide is related to the volcanic rocks. The eruption centres themselves are highly scoriaceous and are therefore very permeable (Ollier, 1988) and so are sites where preferential recharge generates groundwater mounds (e.g. Thompson, 1972; Raiber *et al.*, 2009). Away from the vents, the basalt plains have lower hydraulic conductivity. Prior to the eruption of Mt Rouse and Mt Napier, southwards drainage existed in the region, i.e. the headwaters of the Moyne River (Ollier, 1988). Whereas now, some of this drainage has been diverted to the north. A similar process was described by Thompson (1972) regarding the drainage of the Corangamite region.

The zone where the shallow groundwater flow lines of the Upper Wannon region converge is also where they appear to abut against a wall of basement outcrop: the Otway Group of the Dundas Plateau. If the western margin of basement outcrop is an effective hydraulic barrier, the Upper Wannon region would appear to be a closed sub-basin. However, the more likely scenario is that water from the shallow aquifers continues westwards via channelling into surface water flow and shallow fracture flow in the basement rocks. The ultimate destination of these flow systems is the western side of the Dundas Plateau, as dictacted by the surface water drainage patterns of the Wannon and Glenelg Rivers, i.e. shallow groundwater from the Upper Wannon region is transported via surface water interaction to the recharge zone of the Gambier Sub-basin.

4.3.3 Tyrendarra Sub-basin

The surface drainage of the Tyrendarra Sub-basin consists mostly of southward flowing rivers, such as the Moyne River, the Eumeralla River and the Fitzroy River (Figure 4.5), and some minor westwards flowing rivers, such as the Crawford River which crosses the Dartmoor Ridge and flows into the Gambier Sub-basin. Very few permanent lakes exist in the Tyrendarra Sub-basin, however, many ephemeral swamps are present in the northern area at the margins of individual lava flows (Bennetts, 2005). Condah Swamp, near Lake Condah, was in the past significantly larger, prior to drainage by European settlers (Builth, 2009).

The majority of the eastern half of the Tyrendarra Sub-basin is blanketed by the Pliocene-Quaternary Volcanic Aquifer. The western half contains significant surficial volcanics, but also shows the Plio-Pleistocene Heterogeneous Aquifer, the Upper Tertiary Carbonate Aquifer (Port Campbell Limestone) and the Lower Tertiary Sandy Aquifer outcropping. Both first phase basalt flows and second phase eruption centres are present in the Tyrendarra Sub-basin. The groundwater mounds at several parts of the flowpath represent locally important recharge zones and correspond to outcrops of mainly scoriaceous volcanic rocks (c.f. Figures 4.4 and 7.8).

The shallow groundwater flow system of the Tyrendarra Sub-basin is southwards to the Southern Ocean. It is strongly driven by topographic features; groundwater flow lines originate at hills associated with volcanoes or uplifted basement and converge in the valleys of rivers and creeks. The divisions with its western and northern neighbours have been discussed above. The eastern margin of the watertable catchment of the Tyrendarra Subbasin is a groundwater ridge produced from a topographic high close to the zone of the Warrnambool High basement feature. The topographic ridge is not especially pronounced and could be defined more aptly as the remnant of a plain incised by the subparallel streams: Moyne River to the west and Merri River to the east.

The characteristics of the watertable of the Tyrendarra Sub-basin are also controlled by geological factors. North and east of the 100 mAHD isopotential contour, which is the area where basalt flows are generally thick and laterally contiguous, the isopotentials are smooth and subparallel (Figure 4.4). Conversely, down-gradient from the 100 mAHD contour, the isopotentials become more complex in response to the heterogeneous surficial geology and its related geomorphology. The smooth contours indicate intermediate and regional flow systems are important in the shallow groundwater system. The complex contour patterns indicate that local flow systems dominate in the south.

The topographic relief close to the coast driving shallow groundwater flow in this subbasin supports the strong submarine discharge of groundwater observed in the past (pers. comm. P. Arkell, 2006) and prevents the inland encroachment of the freshwater-saltwater interface. Unlike parts of other sub-basins, the 0 mAHD isopotential contour does not encroach further than the coastline.

4.3.4 Port Campbell Sub-basin

The Port Campbell Sub-basin contains abundant small lakes, mostly resident in volcanic craters in the east. In addition, several creeks and rivers drain south towards the ocean, most notably Hopkins River, which reaches the coast at Warrnambool.

The north of the Sub-basin is dominated by the Pliocene-Quaternary Volcanic Aquifer. Further south the Plio-Pleistocene Heterogeneous Aquifer and the Upper Tertiary Carbonate Aquifer (Port Campbell Limestone) also crop out in more abundance, punctuated by eruption centres. The Plio-Pleistocene Heterogeneous Aquifer is composed of the Hanson Plain Sand in much of this area. In the southeast corner there is significant outcrop of the Upper Tertiary Aquitard and a smaller exposure of the underlying Lower Tertiary Sandy Aquifer.

The watertable in this Sub-basin indicates that shallow groundwater flow is generally to the south and southwest. Groundwater divides with the neighbouring Gellibrand River region and Corangamite region are discussed in Sections 4.3.4.1 and 4.3.5. Flow lines originate in the north where depth to basement is shallow or basement crops out. The north exhibits the smooth isopotentials associated with intermediate flow systems. However, in the central and southern parts of the Sub-basin, complex flow patterns have developed and indicate that local flow systems with short groundwater residence times are dominant. There are two geological factors that are controlling the complex flow patterns. The first is the influence of eruption centres and their volcanic lithofacies with high hydraulic conductivity (Ollier, 1988); the second is the effect of the karst features in exposed carbonate rocks.

The first factor has resulted in the formation of the small, concentric groundwater mounds in the Port Campbell Sub-basin, visible in Figure 4.4. Most associated eruption centres are maars, such as Lake Bullen Merri, and others are scoria cones, such as the small Mount Myrtoon, whose crater is only several hundred metres across. Despite their size, these eruption centres are evidently important local recharge zones. The basaltic eruption centres are not associated with these features (refer to Figure 7.8).

The second factor has produced the cuspate-lobate undulations of the watertable in the southeast of the Port Campbell Sub-basin. These represent complex flow systems that are perpetuated by the karst-incised topography of the exposed carbonates, namely the limestone of the Upper Tertiary Carbonate Aquifer and marl of the Upper Tertiary Aquitard. The pattern in the watertable in this region results in short local flow lines that are not all directed towards the coastline. For example, several small depressions defined by the isopotential contours that are present in the valley of the Curdies River and its tributaries (Figure 4.5) indicate that either shallow groundwater discharges to the surface water features as baseflow or recharges vertically downward into conduits widened by karst processes. Vertical groundwater flow is addressed in more detail in Section 5.3.

Regionally, discharge is ocean bound via coastal seeps and springs (observed in many of the cliffs of the region), via diffuse submarine leakage or via groundwater-surface water interaction. Groundwater discharges to the lower reaches of the rivers as baseflow, evidenced by the way the watertable isopotentials intersect the streams (Figure 4.3). Consequently, discharge to the ocean occurs partly via the estuaries. In particular, the tidal estuary of the Hopkins River has significant interaction with the ocean, and this is likely the reason for the onshore position of the 0 mAHD contour in that region.

4.3.4.1 Gellibrand River region

The Gellibrand River and Kennedy Creek run southwestwards through the centre of this narrow groundwater sub-catchment. Their valleys expose the Lower Tertiary Sandy Aquifer, the Upper Cretaceous Aquifer and the Upper Tertiary and Lower Tertiary Aquitards. The hills to the south east of the valleys are composed of the basement Otway Group and tributaries to the Gellibrand River flow north-west down their flanks. In the hills to the north and west of the river valleys, exposures of the Plio-Pleistocene Heterogeneous Aquifer, the Upper Tertiary Aquitard and the Lower Tertiary Sandy Aquifer and Basement are all present.

The shallow groundwater divide that surrounds the north and west of the Gellibrand River region is related to semi-contiguous topographic highs coincident with the Ferguson Hill Anticline and the Barongarook High (refer to Figures 2.2 and 2.21). The northern groundwater boundary, shared with the Barwon Downs Sub-basin, is not well defined in its own right and is dictated by position of the mound at the Barongarook High.

Shallow groundwater flow in this region generally mimics surface water drainage. Flow converges in the valley of the Gellibrand River and continues to the coast. Interaquifer flow is very important, as demonstrated by the variable surface geology. First, a small groundwater depression in the valley of Kennedy Creek lies within the exposure of the Upper Tertiary Aquitard. This feature is similar to those described above, situated in the karst affected geomorphology of the Port Campbell Sub-basin, and may be linked to deeper leakage with underlying units.

Second, shallow groundwater flow is connected from the northeast of the sub-basin to its south over a structural feature known as the Gellibrand Saddle: the narrowest point between the basement outcrops of the Barongarook High and the Otway Ranges. This is significant because the groundwater north of the Saddle is hosted in the Lower Tertiary Sandy Aquifer and that to the south flows through the Upper Cretaceous Aquifer (Figure 2.1). The connection of these aquifers is possible because the Lower Tertiary Aquitard that usually separates them is locally absent (Leonard *et al.*, 1983). Most previous authors have recognised this connection and have treated the two aquifers as one hydrostratigraphic unit (e.g. Johns, 1968; Thompson, 1972; Blake, 1980; Petrides & Cartwright, 2006). However, the case where the Lower Tertiary Aquitard separates the units is prevalent in the basin. Therefore, in this study which encompasses the entire Otway Basin, the aquifers are treated separately (c.f. Figures 4.14 and 4.21).

4.3.5 Corangamite region

The Corangamite region is almost an entirely internally drained sub-basin, with several northwards and southwards flowing streams terminating in or near the main groundwater discharge feature, Lake Corangamite itself. The closed nature of this catchment is underlined by the great accumulation of dissolved solids in the water of Lake Corangamite. The shallow groundwater flow within this region is hosted in the different lithofacies of the Pliocene-Quaternary Volcanic and Plio-Pleistocene Heterogeneous Aquifers. The former includes low hydraulic conductivity basalt flows and higher hydraulic conductivity scoria and stony rise deposits.

Lake Corangamite was formed after the eruption of the Western District Province volcanics. The first phase of the volcanic eruptions, voluminous lava flows, blocked the extant weak southwards and eastwards drainage system of the Corangamite region and reoriented it to form many lakes in the depressions in lava flows or along the edges of basalt deposits (Currey, 1964; Thompson, 1972; Ollier, 1988). Neotectonic uplift probably assisted with this process. Lacustrine deposits after the first and second phases of volcanism attest to the long history of lakes in the region but the quality, extent and depth of the lakes has undoubtedly fluctuated (refer to Section 3.3).

The groundwater catchment boundaries of this region were formed in much the same way the drainage patterns were. These mechanisms are similar to the two-fold processes described for the Upper Wannon region. Production of groundwater mounds in relation to increased deep drainage at maar and scoria volcanoes results in groundwater divides. In addition, slowing or stopping of old southwards groundwater flow by less permeable lava flows allows for the development of lakes. Both factors are very important for the formation and maintenance of the groundwater flow divides with the Port Campbell Sub-basin and the Barwon Downs Sub-basin.

In contrast to the small groundwater sinks in the karst-affected areas to the south, the depression in the watertable mimicking the shoreline of Lake Corangamite does not imply deep drainage and recharge to the underlying aquifers. There are two important, interdependent reasons for this. The first is the fact that the depression in the Corangamite area is regional in extent because it is at the end of many regional flow lines. Second, at the end of regional flow lines, vertical gradients are theoretically upwards, indicating or causing groundwater discharge (e.g. Tóth, 1963), which is the case for the Corangamite region (Figure 4.4 c.f. Figure 4.13). Leakage from underlying units to the surficial aquifers in this region was also proposed by Cox *et al.* (2007) based on isotopic evidence.

4.3.6 Barwon Downs Sub-basin

The main surface water features of the Barwon Downs Sub-basin are the lakes of the volcanic province in the centre of the region and the northwards and southwards flowing rivers which eventually converge and flow towards the coast in the east.

Shallow groundwater flow in the Sub-basin mimics that of the surface water. Flow begins generally in the highlands to the north or south as meridional and then converges and rotates to flow eastwards. The pronounced groundwater mound at the Barongarook High drives effective recharge and is important for the Lower Tertiary Sandy Aquifer which is exposed there. However, there is a large zone in the centre of the region where hydraulic gradients are very gentle and groundwater flow is accordingly sluggish. This is the region where several lakes have formed at the edges of lava flows or in associated depressions and the lakes act as important pathways for local flow systems, be they dominated by discharge, recharge or flow-through processes (Coram, 1996).

The groundwater divide between the Corangamite region and the Barwon Downs Subbasin is a somewhat weak one. Flow is generally orthogonal to the divide and only a small reduction or increase in hydraulic head in either sub-basin is needed to cause the division to migrate. However, migration of the divide is hindered by the fact that the small watertable mounds at the division's centre are geologically and topographically induced, correlating to the stony rise outcrops and scoria cones in the area of the Red Rock lake complex. Although this area has experienced significant decline in the watertable over the past years (Adler & Lawrence, 2008; Chapter 5), the head level is still relatively higher than neighbouring regions.

4.4 Flow in the Plio-Pleistocene Heterogeneous Aquifer

In the sections on shallow groundwater flow above, a lot has already been presented regarding flow in the Plio-Pleistocene Heterogeneous Aquifer. Hence, it is evident at this stage that significant interaction between the Plio-Pleistocene Heterogeneous Aquifer and other aquifers or surface water bodies is prevalent and extremely important. Some particular points about flow within the hydrostratigraphic unit are explored in further detail below.

One major feature of the Plio-Pleistocene Heterogeneous Aquifer is the dune and swale system that dominates the Gambier Sub-basin. The groundwater flow within this system is complex and local, although exchange with regional and intermediate flow systems also occurs. A representative analogue of the flow system is depicted in Figure 4.2, which shows the flow in aquifers of the Atlantic Coastal Plain and further confirms the similarity between that region and the Otway Basin. Rather than the rivers depicted in Figure 4.2, the depressions of the Gambier Sub-basin are occupied by lagoons, or sometimes by drains, and they can either be recharge, discharge or flow-through features. Their status mainly depends on their position within the regional flow system and their position in relation to the local topography (Winter *et al.*, 1998). As represented in the diagram, the watertable beneath the dune ridges is elevated relative to neighbouring swales (Waterhouse, 1977; White, 1994). This has the important effect of inducing some local flow in the opposing direction to the coastward regional flow.

The dunes of this system are composed of the Bridgewater Formation which is a calcarenite affected by syngenetic karst processes in some areas (White, 1984). The combination of dune and karst landforms together make this an unusual region for groundwater flow. Ephemeral swamps and lakes in sinkholes are known to be conduits for recharge (O'Driscoll, 1960; Herczeg *et al.*, 1997), although flow lines remain local in extent and effective mixing throughout the shallow aquifer does not occur. Mixing is also inhibited by subsurface karst conduits which allow flow of greater velocity when compared to zones of the aquifer where primary porosity is dominant.

At several locations the coastal discharge from the Plio-Pleistocene Heterogeneous Aquifer, particularly the Bridgewater Formation, has been observed in the form of seepage springs, e.g. Cape Otway and Cape Bridgewater (White, 1994). The seeps are not ephemeral but flow rates can vary with the seasons.

Elsewhere in the basin, Pliocene sand, e.g. the Dorodong Sand and the Hanson Plain Sand, is an important host to shallow groundwater flow. As with the Pleistocene carbonates, the flow systems within these aquifers are local and their importance with the intermediate and regional flow systems lies in the hydraulic connectivity with the underlying units (e.g. O'Driscoll, 1960; Love *et al.*, 1993). Whereas the Plio-Pleistocene Heterogeneous Aquifer in the Gambier Sub-basin generally overlies the Upper Tertiary Carbonate Aquifer, in the other regions a more complex relationship is observed. An overlying relationship with the Upper Tertiary Carbonate Aquifer is common. However, in many areas interlayering or contact with the Pliocene-Quaternary Volcanic Aquifer has occurred. Several phases of volcanic and sedimentary rocks can be interfingered in many regions. In addition, there are areas where the Plio-Pleistocene Heterogeneous Aquifer.

Where overlain by the Pliocene-Quaternary Volcanic Aquifer, the Plio-Pleistocene Heterogeneous Aquifer can be confined and will generally receive recharge through the volcanics (e.g. Bennetts, 2005). In this sense, the lithofacies of the Pliocene-Quaternary Volcanic Aquifer and the way they affect groundwater flow are also influential on groundwater flow in the Plio-Pleistocene Heterogeneous Aquifer.

Both the carbonate Pleistocene and siliceous Pliocene parts of the aquifer are utilised
for groundwater resources. Porosity values are about 0.1, although the storage coefficient is less than this where the aquifer is confined, and hydraulic conductivity generally ranges from 1 to 5 m/day, unless karst conduits are influential. Horizontal hydraulic gradients are in the order of 0.002 to 0.0005 in the Gambier Sub-basin but local vertical gradients, encouraged by the topographic highs of the dunes, are likely to be greater. Horizontal hydraulic gradients in the valleys of the Glenelg, Fitzroy, Curdies and Gellibrand Rivers are also higher. Due to the regional scope of this study, attempts have not been made to calculate groundwater velocity or residence time for this aquifer, given its discontinuous nature and the complex, local nature of the flow.

However, estimates of recharge rates have been attempted. Hydrographs of five bores screened in the Plio-Pleistocene Heterogeneous Aquifer and 4 bores screened in the Pliocene-Quaternary Volcanic Aquifer were analysed using the watertable fluctuation (WTF) method of estimating recharge (Figure 4.6). Both total recharge and net recharge were calculated based on seasonal (annual) fluctuations in the hydrographs. Both methods are discussed in detail by Healy and Cook (2002). The WTF results are compared to previous estimates (Table 4.4).

The calculation of total annual recharge using one seasonal scale measurement is not ideal, as the rate of flow to the watertable within the measurement timeframe may be challenged by waterflow away (Healy & Cook, 2002). Notwithstanding this, the approach is still currently employed with success (e.g. Sharda *et al.*, 2006; Cartwright *et al.*, 2007). However, the WTF method cannot be used to calculate recharge from periods where the hydrograph is in recession (Healy & Cook, 2002). In the case of the watertable hydrographs of the Otway Basin, the recession phase reliably begins between October and December and finishes around April or May. The likelihood of significant recharge occurring during these periods is very low, considering that evapotranspiration generally exceeds precipitation in the warmer summer months (Chapter 3). Unfortunately, it is not possible to measure recharge contributions from storm events during the recession phase because of the infrequency of the watertable observations. The total recharge values were derived from an extrapolated antecedent recession curve based on the hydrograph fluctuations. This is a subjective process and creates a degree of uncertainty in addition to those problems mentioned above.

The calculation of net annual recharge using the WTF method involves the relative comparison of consecutive seasonal peaks. The net recharge represents the change in subsurface storage of groundwater (Healy & Cook, 2002), and can be expressed simply as:

$$\Delta S_{aw} = R_{total} + I - O - ET - D \tag{4.5}$$

where ΔS_{gw} is the change in storage; R_{total} is actual recharge to the watertable; I and O are subsurface groundwater influx and outflux; ET is evapotranspiration; D is discharge, e.g. baseflow. Therefore, if the recharge to the watertable is the same volume each year, but other factors differ, the net recharge value may also differ. If net recharge is equal to zero, the water lost from storage has been fully replaced by recharge or influx. The net recharge parameter is useful as a guide to what is happening in the aquifer on a longer time scale. For example, the climate of the region dictates that most of the R_{total} component is incident in the winter, whereas the ET component will mainly be withdrawn from the aquifer over the summer months. Over a period of several years, these fluctuating inputs and outputs may or may not form a balance.

The WTF method requires the application of a suitable specific yield value of the aquifers. Several sources have provided porosity values of 0.1 for Sub-aquifer One of the Plio-Pleistocene Heterogeneous Aquifer (Table 4.1). Given the typical specific yield values of sand and limestone are about 90% of the typical porosity values (Table 4.3), it is reasonable to assume that the specific yield of Sub-aquifer One is close to 0.09. Further, measured values of specific yield for the Bridgewater Formation range from 0.09 to 0.18 (Harris, 1969). Therefore, this range is applied to the hydrograph data for recharge estimation in Sub-aquifer One.

Material	Porosity (%)	Specific Yield (%)
Soil	55	40
Clay	50	2
Sand	25	22
Gravel	20	19
Limestone	20	18
Sandstone	11	6
(semi-consolidated)		
Granite	0.1	0.09
Basalt (young)	11	8

Table 4.3. Typical porosity and specific yield values of different materials (after Heath, 1983).

As for Sub-aquifer Two, porosity and specific yield measurements specific to the Otway Basin are scarce. The clay and silt content of this unit is greater that Sub-aquifer One and, therefore, the porosity is expected to be higher and the specific yield lower (c.f. *Clay* in Table 4.3). Typical specific yield values obtained from sediment dominantly composed of silt, or composed of clay with minor impurities such as sand or silt, range from 0.01 to 0.03 (Cohen, 1963). Therefore, this same range is applied to the WTF method calculations for bores screened in the clay-rich alluvial or lagoonal sediment of Sub-aquifer Two.

Similarly, two different ranges of specific yield have been chosen to apply to the calculations for the lithofacies of the Pliocene-Quaternary Volcanic Aquifer. The first phase basalt flows are given a range of 0.05–0.08 (Thompson, 1972; Heath, 1983). The stony rise deposits and the second phase phreatomagmatic deposits are estimated to have a range of 0.08–0.18. This range is greater than that for the basalt, which is fitting, and the upper limit is defined by the typical value for limestone (Heath, 1983) due to the karst-like nature of the secondary porosity of these deposits (Thompson, 1972).

The hydrographs analysed are presented in Figure 4.6 and the results of the WTF analysis are presented in Figure 4.7; they can be compared to values from other authors collated in Table 4.4. The hydrographs where specifically chosen to represent the different lithofacies of the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers in each region where possible and because they presented behaviour typical in other bores. The number of appropriate monitoring bores was very low in the central part of the basin and hence the Tyrendarra and Port Campbell Sub-basins are presented together. It is evident that the hydraulic characteristics of the aquifer material affect the seasonality of the hydrograph. Apart from the annual fluctuations, it is possible to obtain long term trends from the hydrographs; it can be seen that the trends of these hydrographs are all negative. The rates of long term decline in hydraulic head are displayed in Figure 4.6. This issue is discussed in further detail in Section 5.4.

Existing recharge estimates (Table 4.4) are a useful baseline comparison for the results of this study, both of which are compared to the average total precipitation of each location. The most important parameter for this comparison is the percentage of annual rainfall that annual recharge constitutes. It can be seen from Table 4.4 that typical recharge rates range from 0 to over 20% of rainfall. They vary according to several factors, e.g. the land use and soil types listed in the table, but also according to the annual P/E ratio and the distribution of precipitation and evaporation across the seasonal cycles. A higher rainfall rate and more rainfall in winter are conducive to greater recharge in this region (for comments on variation of climate within the Otway Basin, refer to Chapter 3).

The results for total recharge calculated from the WTF method (Figure 4.7) are too high relative to rainfall in some cases. This is partly because physical hydrogeology methods of estimating recharge tend to yield higher estimates than most chemistry methods (Walker *et al.*, 1990). The other reason is that the upper estimates of specific yield of the aquifers are probably too high. Therefore, when considering the results of the total recharge in Figure 4.7, only the minima are relevant. Accordingly, the ranges of the minima are recorded on the diagram as a percentage of average rainfall. The specific yield may still have been overestimated in some cases, where the proportions of rainfall reach up to more than 50%, in contrast to the lower values in Table 4.4 (most below 20%).

It is evident from the total annual recharge results that lithology is a major factor controlling recharge and this conclusion is supported by others (Table 4.4). The vegetation



Figure 4.6. Hydrographs of bores screened in the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers which were analysed for recharge. Hydrographs are colour coded according to aquifer and labelled with the following information: Bore ID; aquifer lithology type; depth of screen interval (m); and in parentheses, long term rate of change in head (m/yr). These hydrographs are typical of many in the Otway Basin and all happen to show declining (negative) long term trends.

Source	Rech	arge	Proportion o	f Region	Lithology of surficial	Method	Vegetation	Comment
	Min (mm)	(mm)	Precipitation (%)	c	aquifer or soil type			
Holmes & Colville, 1970b		30	5*	N Gambier	overall		overall	
Walker et al., 1990	2	12	$0.3-2^{*}$	N Gambier	clay	chloride and matric suction	pasture	
Walker et al., 1990		80	13^{*}	N Gambier	clay	chloride and matric suction		surface runoff to this point is a significant factor
Stadter, 1989	10	40	$1.7 - 7^{*}$	N Gambier	overall			
Leaney & Herczeg, 1995	20	100	$3.3 - 17^{*}$	N Gambier	sandy	soil chloride mass balance		
Walker et al., 1992	50	70	8.3-12*	N Gambier	dune	chloride and soil moisture mass balances	pasture	
Walker et al., 1992	0		0	N Gambier	dune	chloride and soil moisture mass balances	lucerne or mallee	
Allison & Hughes, 1978		200	29*	S Gambier	sandy			
Holmes & Colville, 1970a	0		0	S Gambier	sandy (UTC)	neutron moisture meter and water balance with rainfall	forest	
Holmes & Colville, 1970b	63	85	10	S Gambier	sandy (UTC) 1	ysimeter and neutron moisture meter	grassland	
Colville & Holmes, 1972	19	73	$2.7 - 10^{*}$	S Gambier	•	estimation based on WTF	forest	
Harrington et al 1999	5		0.7*	Gambier	-	used as a representative value based on other sources		
Allison & Hughes, 1975		27	3.9^{*}	Gambier	ſ	nodellling of tritium concentration in groundwater		lateral inflow is significant
Allison & Hughes, 1975		70	10^{*}	Gambier	-	physical hydrogeology		lateral inflow is significant
Bradley et al., 1995	0	75	$0-11^{*}$	Gambier				recharge varies according to soil type and land use
Raiber & Webb, 2007		50	7	Tyrendarra	eruption centres	chloride mass balance, verified by isotope tracers		
Raiber & Webb, 2007	1		0.2	Tyrendarra	basalt plains o	chloride mass balance, verified by isotope tracers		
Skidmore, 1992			17	Port Campbei	ll overall	alibrated model		
Adler & Lawrence, 2008	120	500	>20	Corangamite	eruption centres	estimation from lake water balances	lower = native vegetati	ion
Adlar & Lawrence 2008	18		್	Corangamite	hasalt nlains	setimation from lake water halances	higher = pasture	
11000 W HUMICHICO, 2000				COL ULIBUILIO	empid ameno		mante research	
Blake, 1980 Lakev & Leonard. 1984			5 14	Barwon Barwon	sandy (LTS) sandv (LTS)		unforested	
Leonard et al., 1983		160	15	Barwon	sandy (LTS)			
Love et al., 1993	78**	149^{**}		Gambier				amount of excess precipitation after evaporation

Table 4.4. Estimates from several methods of recharge in different soil types and types of vegetation in the sub-basins of the Otway Basin. UTC—Upper Tertiary Carbonate Aquifer; LTS—Lower Tertiary Sandy Aquifer; *—values derived from assumed total precipitation (600 mm for N Gambier, 700 mm for Gambier and S Gambier); **—values relate to excess precipitation rather than recharge; sources are listed in full in the bibliography.



Figure 4.7. Total annual recharge and cumulative net recharge to the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers derived from the WTF method. The range in total recharge is due to the range of specific yield applied to the calculation. The average precipitation values are from data for the years 1961–1990 and are representative of the locations of the bores (Figure 3.4). The range in minimum recharge as percentage of rainfall is recorded next to the bore numbers. The net recharge is presented as cumulative totals since 1992. Readings were too infrequent to create a cumulative plot for a longer time period.

type at each bore location is not recorded in this study, however, this is known to have a strong influence on the amount of recharge. Some studies have shown that almost no recharge reaches the watertable in the basin directly below certain types of vegetation (Holmes & Colville, 1970a). These factors account for much of the variability in the recharge estimates of this study. In addition, there is the issue of variability in precipitation and evaporation. This is somewhat avoided by comparing recharge values as a proportion of total local annual rainfall.

One final factor that may be contributing to the overestimation of recharge at some sites is lateral flow, both surficial and subsurface. Runoff can redistribute rainfall, allowing it to be recharged in spatially disproportionate amounts. Such surface drainage can create ephemeral swamps that may act as temporary recharge features (Walker *et al.*, 1990). Generally such recharge features contain fresh water, as distinct from other lagoons and lakes (e.g. Bool Lagoon; O'Driscoll, 1960) which act as local- or intermediatescale discharge zones and where evapotranspiration has concentrated the salt in the water. Similarly, subsurface lateral inflow, originally recharged elsewhere, can contribute to the seasonal rise in the hydrograph and would be interpreted as *in situ* recharge by the WTF method. Although the issue of lateral flow presents a significant challenge to estimating recharge, it underlines the importance of preferential recharge zones, be they related to topography, soil or aquifer type or vegetation cover, to the sustainability of the groundwater resource.

Fortunately, all of the uncertainty surrounding the above factors is eliminated when the water balance is considered, as in the form ΔS_{gw} . If the inputs, including variable rainfall, and outputs, including changes in transpiration because of different vegetation, are balanced, the change in storage will be close to zero. As can be seen in Figure 4.6, the overall trends in the hydrographs is negative. It therefore follows that the majority of the annual net recharge values are negative and the cumulative total of them is also negative (Figure 4.7). The implication of this finding is that there is a declining trend in the recharge or lateral influx, and/or an increasing trend in evapotranspiration, discharge or outflux (Equation 4.5).

Although no declining or increasing trend in seasonal recharge is discernible from the total annual recharge analysis in this study, a more comprehensive analysis may be required to rule this out as the cause of the decline in storage. If recharge is not steadily declining, it follows that the P/E ratio over the important winter recharge months must be relatively unchanging. The WTF results appear to demonstrate that an increase in outflows is the most likely cause. The recovery of the storage value in the more conductive aquifers in 2000 and 2004 indicates that the trend is not irreversible (Figure 4.7). However, the less conductive aquifers are more indicative of the long term trends because flow is slower within them and hence lags are greater. The question of why storage is declining is addressed further in Section 5.4.

4.5 Flow in the Pliocene-Quaternary Volcanic Aquifer

Several aspects of groundwater flow in the Pliocene-Quaternary Volcanic Aquifer have been addressed thus far, including: its dual porosity and the confining nature of its clay layers (Section 4.1.3); the role of the permeable lithofacies of its eruption centres forming important recharge zones, the local nature of flow and the interaction with surface water, e.g. the changes wrought upon the drainage system by the first phase of volcanism (Section 4.3); and its recharge rates and patterns, e.g. recharge varying according to lithofacies and a long term decline in shallow groundwater storage (Section 4.4).

As with the Plio-Pleistocene Heterogeneous Aquifer, the local nature of the flow systems within the Pliocene-Quaternary Volcanic Aquifer and the regional scale of this study prevents detailed estimates of groundwater velocity or residence times. However, should such detailed analyses arise in future studies, appropriate hydraulic characteristics will be necessary. The lithofacies of the Pliocene-Quaternary Volcanic Aquifer dictate its hydraulic characteristics and hence values derived from single pump tests are highly variable. For example, Dahlhaus *et al.* (2002) reported that individual values of hydraulic conductivity for different lithofacies within the Pliocene-Quaternary Volcanic Aquifer can range from 0.0008 to 86 m/day. This range is marginally broader than that presented in Table 4.1 (0.4–48 m/day). Bennetts (2005) avoided the complexity in characterising the aquifer by assigning a hydraulic conductivity value representative of most fractured rock aquifers, 0.43 m/day, and an effective porosity of 5%, although Thompson (1972) found some basalt and ash to have a porosity greater than 5%. Bennetts' method is satisfactory, although it would be desirable to represent areas of changing hydraulic conductivity.

The Pliocene-Quaternary Volcanic Aquifer is extensively used as a water resource in the Otway Basin. Most applications are for stock and domestic purposes, but some larger demands on the resource also exist, such as town supply, dairy wash and irrigation. The aquifer is variably appropriate for such applications depending on the groundwater quality and the local typical yield, both of which are often linked to the aquifer material type. For example, as previously mentioned, Thompson (1972) compared the pervasive secondary porosity of the stony rises (being the abundant fractures and joints), to that of a karst aquifer, and proposed that groundwater flow and recharge in the stony rises is analogous to karst-affected flow. These areas typically provide good quality water because of the effective drainage preventing evapotranspiration and also high volume yields because of the high hydraulic conductivity. Perched watertables are often observed in the Pliocene-Quaternary Volcanic Aquifer (O'Driscoll, 1960), possibly to do with the layering of units within the aquifer system of variable hydraulic conductivity. These bodies of water are only small and are not reliable for water supply. They are also presumably ephemeral, perhaps similar to those observed in sandy soil profiles overlying the Upper Tertiary Carbonate Aquifer associated with failed wetting fronts after rain events (Holmes & Colville, 1970a).

In recent years, a number of the lakes in the volcanic province have dried out entirely or have experienced significant lowering of lake levels (e.g. Adler & Lawrence, 2008). These lakes are innately linked to groundwater, despite the fact that they may not be discharge features. Therefore, although the main portions of the lake budgets, volumetrically speaking, are input from rainfall and outflux via evaporation (Coram, 1996), the lake levels are in some way dependent upon the position of the watertable. The ephemeral nature of the perched watertable occurrences are a testament to this dependency. It is no coincidence that the recent drying of lakes comes with a long term decline in hydraulic head levels. However, the questions that remain to be answered are: Is the fall in lake levels being driven by loss of head, and, if so, what is causing the fall in the watertable? or alternatively: Are both the fall in lake level and head level occurring simultaneously, both driven by another factor? These issues are discussed in greater detail in Section 5.4.

4.6 Flow in Upper Tertiary Carbonate Aquifer

The upper part of the Upper Tertiary Carbonate Aquifer is laterally contiguous throughout much of the basin, except in the region of the Dartmoor Ridge, where the Gambier Subbasin host rock, the Gambier Limestone, is distinguished from its eastern counterpart, the Port Campbell Limestone. The aquifer reaches a thickness of greater than 300 m onshore, containing groundwater of generally high quality. The flow system within the Upper Tertiary Carbonate Aquifer (Figure 4.8) bears similarities to the watertable elevation (Figure 4.4) because the aquifer is exposed at the surface or lies close to the surface in much of the region. Where it is exposed, the shallow portions of the aquifer are subject to the heterogenous flow patterns of local systems (outcrop is marked in Figure 4.8). However, flow in this aquifer is mostly on an intermediate and regional scale and isopotentials of the deepest levels of the aquifer would be considerably smoother than those in the map. Where the Upper Tertiary Carbonate Aquifer is blanketed by the Plio-Pleistocene Heterogeneous or Pliocene-Quaternary Volcanic Aquifers, the intermediate and regional flow systems within it interact with the local and intermediate flow systems of the shallower units.

The Clifton Formation hydraulic head data are not included in the map (Figure 4.8) because of the vertical gradients that often exist between it and the other components of





the Upper Tertiary Carbonate Aquifer represented in the map: the Gambier Limestone and the Port Campbell Limestone. It was not possible to construct a separate potentiometric map for the Clifton Formation due to the paucity of data points in many regions of the basin. Some aspects of groundwater flow in the Clifton Formation are discussed further in Section 5.3. One of the important recharge zones for the Clifton Formation is the Upper Wannon region, where the Upper Tertiary Carbonate Aquifer is close to the surface or is exposed in river valleys (Leonard, 2003).

Both limestone units of the Upper Tertiary Carbonate Aquifer have been characterised by localised karstification which affects flow throughout the basin. This secondary porosity is an important part of the flow system and locally increases specific capacity and hydraulic conductivity dramatically (Shugg, 1981). However, the extremely localised effect of the karstic porosity on these parameters is an indication that it is not interconnected, and that the main mechanism for regional flow is through primary intergranular porosity. Dominance of intergranular flow on the regional scale is supported by Love *et al.* (1993). One particular example of flow in karst fractures being local or disconnected is Blue Lake, a crater lake at Mt Gambier. The crater lake acts as a window onto the Gambier Limestone and the water of the lake supplies the town of Mt Gambier. The water is known to flow in and out of the lake via both primary and secondary porosity. Despite this, the lake water has remained almost untouched by various forms of potential pollution: unregulated disposal of waste and urban storm water drainage in nearby sinkholes (Vanderzalm & Dillon, 2006). The lack of detection of pollution thus far indicates that advection is slow enough to represent transport through primary porosity, and that the karst features are not regionally connected. The important implication of the dual porosity flow is that karst affected areas will act to locally increase flux volumes in response to higher hydraulic conductivity.

Due to similarity of the orientation of the flow systems depicted in the maps of Figures 4.4 and 4.8, the reader is referred to Section 4.3 for a detailed flow direction analysis. The following points are made in this section because they describe the aspects of the groundwater flow in the Upper Tertiary Carbonate Aquifer where it differs significantly from the shallow groundwater flow system.

The recharge zones for the regional flow system within the Upper Tertiary Carbonate Aquifer are inland (Figure 4.8). However, recharge to the local flow systems of the Upper Tertiary Carbonate Aquifer occurs throughout the entire area, provided vertical gradients allow it. In the west, this occurs particularly where karst features facilitate rapid percolation (Sprigg, 1952; O'Driscoll, 1960) and is evidenced by the lack of surface drainage, despite the reliable rainfall. Leaney and Herczeg (1995) concluded that point source recharge from karst features is about 10% of the total recharge in the Gambier Sub-basin. The process of karst conduit flow also applies in the east. In addition, volcanic eruption centres are linked to local- or intermediate-scale recharge zones within the Upper Tertiary Carbonate Aquifer.

A naturally occurring potentiometric pressure low is known to have persisted near the town of Penola (Figure 4.5) since at least 1965 (Colville & Holmes, 1972; Love *et al*, 1993), albeit with differing morphology and extent over the years (Figure 4.8). This area of lower head creates a strong downwards hydraulic gradient between the Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers (c.f. Figure 4.4) and reflects a zone of preferential downwards flux aided by secondary karst porosity. Although not previously documented, a similar depression exists north of the Surrey River, near Heywood. This area is most likely a similar zone of downward infiltration.

In the inland regional recharge regions, the vertical hydraulic gradient between the watertable and the Upper Tertiary Carbonate Aquifer is generally downwards, allowing diffuse drainage in addition to the karst-aided flow already described.

However, north of Dartmoor, close to the Glenelg River, the shallow vertical gradient is locally upwards. This is due to the watertable having been lowered in the region of the river's deep incision, whereas pressure at depth is less affected. The result is that the river receives baseflow from the Upper Tertiary Carbonate Aquifer, even in the regional recharge zone of the aquifer. Baseflow to the Glenelg has been determined to be a greater proportion of the River's flow in successively lower reaches. Baseflow has been estimated using analysis of tritium concentration in the river water and groundwater as approximately 180 L/sec in the reach north of Dartmoor, compared to inflow of 110 L/sec over the same reach from tributaries (Allison *et al.*, 1975).

Baseflow seepage and springflow to the Glenelg River is known to occur currently and in the past due to evidence gleaned from caves at Drik Drik, just south of Dartmoor, and also in the lower reaches. The caves at Drik Drik were partly formed by vadose speleogenesis around 400 ka (White, 2005). For some time since their formation they have been partially or fully saturated and act as subterranean tributaries to the Glenelg River (White, 2005).

Regional discharge zones of the aquifer are generally linked to the coast, either via coastal springs or diffuse submarine discharge. Discharge from local flow systems within the Upper Tertiary Carbonate Aquifer can occur at other points along the regional flow line, e.g. the springs associated with the Tartwaup Fault (Sprigg, 1952; O'Driscoll, 1960; Figure 2.21). Baseflow to surface water features is another way that localised discharge can occur. The lower reaches of the creeks and rivers of the Port Campbell and Tyrendarra Sub-basins are good examples of this. In addition, locally important upwards gradients from the Upper Tertiary Carbonate Aquifer exist where it shares a contact boundary with the Pliocene-Quaternary Volcanic Aquifer at the surface, i.e. usually at the edge of basalt flows (e.g. south of Mt Emu Creek). Alternatively, local discharge zones from the Upper Tertiary Carbonate Aquifer may occur via flow into the overlying formations.

The regional discharge zones are more voluminous, e.g. the coastal karst springs of the Gambier Sub-basin. Such springs at Eight Mile Creek, Deep Creek and Piccaninnie Ponds together discharge approximately 110,000 ML/year to the ocean (Stadter & Yan, 2000). As with the discharge springs of the Plio-Pleistocene Heterogeneous Aquifer (e.g. those at Pt Flinders; Shugg, 1984) karst flow is an important feature of the formation of these features. Some of these regional discharge springs demonstrate seasonal fluctuations (e.g. Ewens Ponds; Sprigg, 1952). In addition, submarine springs have been recorded in the Otway Basin. Submarine springs exist in coincidence with faults at Cape Nelson and Port Macdonnell (Figure 4.5; Sprigg, 1952; Boutakoff, 1963). The discharge volume of these features was so great that fishermen have been known to replenish their water supplies from the surface of the ocean, approximately 2 miles offshore from Cape Nelson (Sprigg, 1952; Pers. comm. P. Arkell, 2005).

The location of these springs, onshore and in submarine areas, flags the general position of regional discharge from this aquifer and coincides with the area just up-gradient of the freshwater-saltwater interface. Discharge occurs at the interface because groundwater flowing oceanward is diverted upwards by a wedge of density-driven seawater at the bottom of the aquifer. The theoretical morphology of the interface in an unconfined aquifer and the associated flow lines are presented schematically in Figures 4.9 and 4.10. As the Upper Tertiary Carbonate Aquifer is generally unconfined at the coast, the location of its interface can be estimated by the Ghyben-Herzberg equation:

$$\rho_s \cdot g \cdot z = \rho_f \cdot g(h_f + z) \tag{4.6}$$

which is an analytical solution that assumes hydrostatic conditions, where ρ_s and ρ_f are the seawater and freshwater densities, g is the acceleration due to gravity, z is the depth below sea level of the interface and h_f is the height of the watertable above the sea level (see Figure 4.9 for relationships).

While this early equation is useful, subsequent research confirmed the nature of the interface to be hydrodynamic, rather than hydrostatic (Hubbert, 1940; 1969). Still more recent investigations have proven that the interface is an interactive zone of mixing and cycling between the two water bodies, resulting in submarine groundwater discharge (SGD) which contains some component of recycled marine water (e.g. Freeze & Cherry, 1979; Domenico & Schwartz, 1997; Smith, 2004; SCOR & LOICZ, 2004; Zektser & Dzhamalov, 2007). The cyclic flow of SGD and some of its driving mechanisms are depicted schemat-



Figure 4.9. Diagramatic representation of the theoretical sharp interface between groundwater in an unconfined aquifer and ocean water showing the parameters of the Ghyben-Herzberg solution (after Bear *et al.*, 1999).



Figure 4.10. A schematic diagram of interaction of sea water and groundwater, showing how submarine porewater exchange occurs in confined and unconfined aquifers and aquitards and some of the driving mechanisms (after Taniguchi *et al.*, 2002).

ically in Figure 4.10 (Taniguchi *et al.*, 2002). The mixing of the two water types around the interface, creating a diffuse zone rather than a sharp interface, renders the Ghyben-Herzberg solution unrepresentative in another sense. Notwithstanding these shortcomings, it is still a useful analytical tool when numerical modelling is not possible. Finite difference modelling of fluid flow and solute transport is more desirable (e.g. Pinder & Cooper, 1970; Khublaryan *et al.*, 1996; Werner & Gallagher, 2006), but was outside the scope of this study.

Equation 4.6 shows that the position of the freshwater-saltwater interface is a function of the densities of the water bodies and the freshwater hydraulic head relative to sea



Figure 4.11. Diagramatic representation of the parameters used to calculate the position of the toe of the freshwater-saltwater interface at the relatively impermeable base of an unconfined aquifer (after Werner & Simmons, 2009).

level, i.e. the hydraulic gradient. However, it is also dependent on other parameters, such as the discharge flux of the aquifer, which in turn is linked to the aquifer's recharge and flow-through flux. In addition, parameters that affect flux, like aquifer thickness and hydraulic conductivity are also important. These relationships are visible through alternative analytical solutions for the interface in unconfined aquifers which define the horizontal distance from the coastline to the toe of the interface $(x_T; Figure 4.11)$. One such equation is (Domenico & Schwartz, 1997):

$$x_T = \frac{K z_0^2}{\alpha q_0} \tag{4.7}$$

and another is (Custodio, 1987; as quoted by Werner & Simmons, 2009):

$$x_T = \frac{q_0}{W} - \sqrt{\frac{q_0^2}{W^2} - \frac{K(1+\alpha)Z_0^2}{W\alpha^2}}$$
(4.8)

where

$$\alpha = \frac{\rho_f}{\rho_s - \rho_f} \tag{4.9}$$

and where z_0 is the aquifer thickness below sea level [m] (constant because aquifer bottom is assumed to be horizontal); q_0 is the discharge flux per unit length of coastline [m²/day]; W is the recharge [m/day]. The relationships of these parameters is shown in Figure 4.11.

These analytical solutions are applied where possible to three instances of an onshore presence of the freshwater-saltwater interface within the Upper Tertiary Carbonate Aquifer. The aim is to compare the theoretical results with the observed location of the



Figure 4.12. Electrical conductivity time series for groundwater from three nested bores at Ewens Ponds, near Port Macdonnell. The bores' screen intervals are provided as elevations in mAHD.

interface in each case. The three locations used in this study, one of which has been studied previously (King & Dodds, 2002), are near the coastal towns of Port Macdonnell, Port Fairy and Warrnambool. They were identified because of the presence of high salinity groundwater at the coast (EC > 30,000 μ S/cm) surrounded by fresh groundwater (EC \approx 1000–2000 μ S/cm). The features can be seen in Figure 6.8 (Chapter 6).

The possibility of sea water intrusion in the Port Macdonnell region was previously identified through land resistivity surveys conducted in 2002 by the Department of Water, Land and Biodiversity Conservation (Kind & Dodds, 2002). The surveys, which were short landward transects from the coast at 5 locations, revealed low resistivity areas in the aquifer at depth, indicating the possible presence of high salinity water. As a result of that initial investigation, new observation bores were drilled at various depths and monitored for salinity by the Department. One location, near Ewens Ponds (Figure 4.5) and approximately 1.1 km from the coastline, has three bores screened in the aquifer. In the last few years, the deepest of these has recorded a dramatic increase in the groundwater electrical conductivity (EC), whereas the EC in the shallower levels has remained low and very stable (Figure 4.12). This effectively shows that the diffuse freshwater-saltwater interface has migrated over the screen interval of the deepest bore (CAR061) and that the mixing zone around the interface is yet to affect the other bores. Consequently, the current vertical position of the interface at this location, i.e. 1.1 km from the coast, is between -120 and -150 mAHD.

The Ghyben-Herzberg solution was solved for this case using the water densities and

hydraulic head values from these bores. The theoretical results indicate that the interface should lie around -83.1 or -62.3 mAHD, which is notably shallower than the actual position. Equation 4.7 was also applied, and produced estimates of x_T around 2–5 km, depending on the magnitude of the hydraulic gradient. These latter values cannot be confirmed or refuted because bores penetrating to the base of the aquifer are not present at the necessary locations. Some of the discrepancy between the observed and theoretical interface positions for this site may be attributed to the assumptions implicit in Equation 4.6.

However, the discrepancy also indicates that the interface migration is lagging behind the changes in hydraulic boundary conditions (Bush & Lawrence, 2006). Sea level changes or changes in hydraulic head affect the balance between the freshwater and seawater, and, in response, migration towards a new equilibrium position can occur. In the case of Ewens Ponds, the landward migration of the interface has been observed (Figure 4.12) and, if the calculations are representative, further migration to a shallower equilibrium is anticipated. The issues of concern are what has caused the instability in boundary conditions that forced the interface migration and why is the shift currently observable at this site.

First, the cause of the landward interface migration is either owing to a sea level rise or a fall in onshore pressure head. The latter has been observed in unconfined aquifers connected to the Upper Tertiary Carbonate Aquifer (Figure 4.6). This decline may be a response to groundwater abstraction or climate change and is discussed in detail in Chapter 5.4. In discerning the main causal factor in the current freshwater-saltwater interface migration, the lag time involved in the aquifer response to the different boundary condition changes is important.

It is known that the lag time between onset of disturbances at a certain location to steady state flow and the observation of resultant changes at a different location is related to the distance between the locations (Balleau, 1988). The maximum impact of distal disturbances may be felt long after the disturbance (e.g. pumping) has ceased and a new equilibrium could take over a millennium to develop, whereas the effects of proximal stresses are felt more quickly (Bredehoeft & Durbin, 2008). In the case of Ewens Ponds, the observation location is the zone of the interface. Therefore, sea level and associated coastal pressure changes are proximal and recharge zone processes and extraction (e.g. at Mt Gambier c. 30 km inland) are distal. This means that only recent sea level changes, e.g. the last 50 years, need be considered as currently influential; but that inland pressure head decline further in the past, e.g. the last 200 years, should be analysed. The rate of sea level rise has on average been 0.17 m/100 yrs for the last 100 years (IPCC, 2007). The recorded decline in hydraulic head of the Upper Tertiary Carbonate Aquifer is commonly in excess of 5 m/100 years (0.05 m/yr; Section 5.4.2), it is clear that climate change and/or

groundwater extraction is the cause of this seawater intrusion, rather than sea level change, and this shift in the interface is observed now, rather than at the onset of the stresses in the past, due to the lag explained above.

Second, the issue of why the shift is observed in certain locations, including Ewens Ponds, is related to aquifer diffusivity. This parameter also affects the lag time in the aquifer's response to changes in boundary conditions (Balleau, 1988). The aquifer diffusivity, equal to the ratio of aquifer transmissivity to storage coefficient, is linearly related to the time lag (Sophocleus, 2002; Parkin *et al.*, 2007). According to the parameters of Table 4.1, and the thickness of the aquifer in the region of Ewens Ponds (Figure 2.15), the diffusivity there is about 4,700–180,000 m²/day for non-karstic areas (c.f. low/moderate diffusivity sandstone: 2,000–10,000 m²/day; Parkin *et al.*, 2007) and could be 74,300–405,000 m²/day for karst-affected rock (c.f. high diffusivity chalk: 100,0000–400,000 m²/day; Parkin *et al.*, 2007). The higher aquifer diffusivity of karst makes it more susceptible to seawater intrusion sooner than other areas and this is indeed the case of the Ewens Ponds area, with the Ponds themselves being major karst springs. In fact, the saline intrusion has targeted the Camelback Member of the Gambier Limestone, a highly fractured and karstic unit which is preferentially tapped for irrigation supply (Section 2.3.7).

Unfortunately, less information exists regarding the Victorian locations of identified salt water in coastal areas, Port Fairy and Warrnambool. No time series data of salinity are available for the relevant boreholes and so the timing of the movement of the interface to its current position cannot be constrained. These locations coincide with areas where the hydraulic head in the aquifer has receded to 0 mAHD or deeper (Figure 4.8). Head values below mean sea level are periodically maintained in this aquifer by pumping and increase the risk of seawater intrusion. According to Equation 4.6, a freshwater head level of 0 mAHD indicates that the interface should be present at sea level in those near-coast inland areas. Yet, despite saltwater in a few coastal boreholes, many areas with such low head still host fresh groundwater (EC = $1,00-2,000 \ \mu S/cm$). This demonstrates a similar lag to that described above and indicates movement of the interface in the future should be anticipated regardless of future boundary condition changes (Bush & Lawrence, 2006).

The Port Fairy site is only defined by one salinity reading in 1990 from one bore. This reading is reliable, however, no further evidence exists and the bore is privately owned so could not be sampled for oxygen or hydrogen isotope analysis in this study. The bore, number 48818, is located about 160 m from the coast and the recorded chloride and sodium concentrations in the water in 1990 were 13,000 and 6,800 mg/L respectively. The TDS is therefore significantly lower than seawater, meaning that the bore was tapping the mixing zone and the interface lay beneath the screen interval elevation (-30.1 - 35.8 mAHD). For comparison, theoretical solutions of the depth of the interface for hydraulic head values

further inland (1,200 m from coast) are -16--95 mAHD. The lag in the interface shift is represented in these data as no high salinity readings are present in those depth intervals.

It is difficult to compare the measured depth of the interface at Warrnambool with a theoretical depth because the actual current depth cannot be accurately measured using the existing data. The gradient from fresh to saline water is obscured by the fact that saline water is present both close to the surface, and at depth (Figure 6.6 c.f. Figure 6.8) the surface, too. This indicates that the brackish water of the Hopkins River Estuary may be influential in the process of groundwater salinisation. Also important to the interface equilibrium issue around Warrnambool is the fact that the town's supply comes from the Upper Tertiary Carbonate Aquifer. Lastly, areas of high aquifer diffusivity are likely to be present in the region because of the presence of karst discharge springs (refer to cave and karst feature density diagram in Figure 7.8). This could be the reason the rate of sea water intrusion is not laterally consistent across the coastline.

In summary, the discharge from the Upper Tertiary Carbonate Aquifer occurs at the freshwater-saltwater interface, either near the coast or in the shallow submarine zone via convective and hydraulic flow mechanisms listed in Figure 4.10. Additionally, some coastal discharge occurs as evapotranspiration, potentially through surface water bodies such as lagoons (Winter et al. 1998; SCOR & LOICZ, 2004; Evans, 2007). The interface position and depth with respect to the coastline and sea level is sensitive to aquifer heterogeneity and groundwater use. Further, recent changes to the hydraulic conditions onshore are causing a landward shift in the interface that can be expected to continue into the future until a new equilibrium is reached.

4.6.1 Groundwater velocity and residency in the Upper Tertiary Carbonate Aquifer

As part of this study some indicative calculations of regional groundwater velocity and residence times in the Upper Tertiary Carbonate Aquifer have been made, assuming steady state flow and applying Darcy's Law (Hubbert, 1969). First, the Darcy flux, or Darcian velocity, is calculated by:

$$q = K \cdot \frac{dh}{dl} \tag{4.10}$$

where q is Darcy flux; K is hydraulic conductivity standardised to 20°C; $\frac{dh}{dl}$ is the horizontal hydraulic gradient. Then the groundwater velocity is calculated by:

$$v = \frac{q}{n_e} \tag{4.11}$$

where v is groundwater velocity; and n_e is effective porosity. Velocity is then used to

calculate a residence time for a groundwater flow path of specific length. In this study, each of these steps is completed for several representative values of the input parameters to gain a range of velocities and residence times (Table 4.5).

The locations of the analyses are flow paths in the Gambier, Tyrendarra and the Port Campbell Sub-basins, which are aligned with the cross sections presented in Chapters 2 and 5. Ranges of hydraulic conductivity and porosity measured in the aquifer in each sub-basin were taken from the data in Table 4.1. Four horizontal hydraulic gradients were calculated for each flow path to characterise the variability present. The gradients were calculated from pairs of bores that tap the Upper Tertiary Carbonate Aquifer; the difference between their density-corrected head values was divided by the horizontal distance between them. The pairs of bores were chosen for their position near the hinge-line of each of the flow paths, i.e. the mid-point of the regional flow system where vertical gradients are weakest and where the flow is dominated by the horizontal hydraulic gradient. The bore pairs were also chosen to avoid areas of intensive abstraction. Ideally, the vertical and horizontal components of flow should be separated completely (Post *et al.*, 2007), but this was not possible given the data available.

The results in Table 4.5 indicate that groundwater velocity in the Upper Tertiary Carbonate Aquifer is fastest in the Gambier Sub-basin, despite the presence of lower hydraulic gradients, due to the high hydraulic conductivity associated with karst features. Port Campbell Sub-basin groundwater has velocity estimates at are similar with less variation, whereas the Tyrendarra Sub-basin has a very wide range of residency estimates. The validity of these calculated ranges was assessed through comparison with radiocarbon ages of groundwater at Nangwarry (Gambier Sub-basin) and near Hawkesdale (Tyrendarra Sub-basin) (town locations in Figure 4.5).

The groundwater at Nangwarry was probably recharged close to 1,600 years ago and is around 20 km from the possible recharge zone (Love *et al.*, 1994). However, the nature of the flowpath in that region is such that some local mixing of modern recharge is likely to have affected the ¹⁴C age. Most of the maximum hydraulic residence times for a flow path of 20 km are commensurate with the ¹⁴C age; whereas the hydraulic minima are far too short, even if some mixing has occurred (Table 4.5). This indicates that the maximum hydraulic conductivity used in the estimates (60 m/day) is not representative of the regional flow system and is only relevant on the local, karstic scale at which it was measured. Therefore, although derived from a crude model, the hydraulic calculations confirm that the majority of the flow in the aquifer is transported via intergranular flow, a model in keeping with that described by Leaney and Herczeg (1995).

West of Hawkesdale, about 35 km from the recharge zone in the Tyrendarra Sub-basin, groundwater residence time was measured by radiocarbon to be 7,550 years (Bennetts,

4.5. Hydraulic groundwater velocity and residence times in the Upper Tertiary Carbonate Aquifer. dh/dl —hydraulic gradient; K—horizontal hydraulic trivity (standardised to 20°C); n_e —effective porosity; q—Darcy flux (Equation 4.10); v—hydraulic groundwater velocity (Equation 4.11). Distances were ated from Figure 4.8. Radiocarbon ages are from Love <i>et al.</i> , (1994; Nangwarry) and Bennetts (2005; Hawkesdale). Locations are shown on a map in Figure	asin Horizontal K(max) K(min) n _e (min) n _e (max) q(max) q(min) v(max) v(min) Location Distance to Residence Residence 14C Age	
Table 4.conductivestimated4.5.	Sub-basir	

Sub-basin F	Horizontal	K (max)	K(min)	n e(min) n	le(max)	q(max)	q(min) v	r(max)	v(min)	Location	Distance to	Residence	Residence	e ¹⁴ C Age
0 4	lh/dl at ningeline									for comparison	location from recharge zone	time at location	time at location	(corrected)
												(max)	(min)	
		[m/day]	[m/day]]	[m/day]	[m/day]	m/day] [m/day]		[km]	[years]	[years]	[¹⁴ C years]
Tyrendarra	0.0026	1.5	0.1	0.05	0.2	0.004	0.0003	0.08	0.001	15 km W of Hawkesdale	35	72,949	1,216	7,550
Tyrendarra	0.0031	1.5	0.1	0.05	0.2	0.005	0.0003	0.09	0.002	15 km W of Hawkesdale	35	61, 387	1,023	7,550
Tyrendarra	0.0039	1.5	0.1	0.05	0.2	0.006	0.0004	0.12	0.002	15 km W of Hawkesdale	35	49,556	826	7,550
Tyrendarra	0.0077	1.5	0.1	0.05	0.2	0.012	0.0008	0.23	0.004	15 km W of Hawkesdale	35	24,944	416	7,550
Gambier	0.0012	60	7	0.2	0.4	0.072	0.008	0.36	0.02	Nangwarry	20	2,616	153	1,600
Gambier	0.0020	09	7	0.2	0.4	0.122	0.014	0.61	0.04	Nangwarry	20	1,545	90	1,600
Gambier	0.0031	60	7	0.2	0.4	0.185	0.022	0.93	0.05	Nangwarry	20	1,014	59	1,600
Gambier	0.0035	60	7	0.2	0.4	0.208	0.024	1.04	0.06	Nangwarry	20	903	53	1,600
Pt Campbell	0.0034	7.2	4.8	0.2	0.4	0.024	0.016	0.12	0.04	c.f. Tyrendarra flow path	35	2,353	784	
Pt Campbell	0.0038	7.2	4.8	0.2	0.4	0.027	0.018	0.14	0.05	c.f. Tyrendarra flow path	35	2,099	700	
Pt Campbell	0.0042	7.2	4.8	0.2	0.4	0.030	0.020	0.15	0.05	c.f. Tyrendarra flow path	35	1,906	635	
Pt Campbell	0.0047	7.2	4.8	0.2	0.4	0.034	0.023	0.17	0.06	c.f. Tyrendarra flow path	35	1,700	567	
Pt Campbell	0.0034	7.2	4.8	0.2	0.4	0.024	0.016	0.12	0.04	c.f. Gambier flow path	20	1,345	448	
Pt Campbell	0.0038	7.2	4.8	0.2	0.4	0.027	0.018	0.14	0.05	c.f. Gambier flow path	20	1,200	400	
Pt Campbell	0.0042	7.2	4.8	0.2	0.4	0.030	0.020	0.15	0.05	c.f. Gambier flow path	20	1,089	363	
Pt Campbell	0.0047	7.2	4.8	0.2	0.4	0.034	0.023	0.17	0.06	c.f. Gambier flow path	20	971	324	

2005). The hydraulic residence times for a flow path of the same length range from 416 to 72,949 years (Table 4.5). This indicates that only the mid-range hydraulic velocities are representative of the groundwater velocity at a regional scale. The upper and lower extremes of the velocity estimates are due to low porosity (5%) and low hydraulic conductivity (0.1 m/day) inputs respectively. Therefore, it is probable that these measured values are not appropriate for upscaling to represent the conditions for regional flow. A better understanding of the variation of the hydraulic characteristics of the Upper Tertiary Carbonate Aquifer in this sub-basin is required before more constrained estimates can be made.

The Port Campbell Sub-basin calculations reveal residence times that are similar to those of the Gambier Sub-basin and shorter than those of the Tyrendarra Sub-basin for flow paths of appropriate length (Table 4.5). No radiocarbon analyses are currently available for the Upper Tertiary Carbonate Aquifer in the Port Campbell Sub-basin, so a direct validation of residence times was not possible. This sub-basin has generally higher hydraulic gradients that the other regions, a result of the more pronounced topographic relief of this region.

Thus, the range of hydraulic characteristics and gradients in each of the sub-basins results in differing groundwater velocity and residence times. This has an impact on groundwater quality for several reasons. First, it can be concluded that the climate and surface conditions over at least the last 10,000 years have had an effect on the recharge conditions for the aquifer and are therefore relevant to the current water composition. Second, the rate of residence time in the aquifer, balanced with kinetic and thermodynamic controls on solubility, dictates the degree of water-rock interaction that the groundwater experiences.

4.7 Flow in the Upper Tertiary Aquitard

Groundwater flow in the Upper Tertiary Aquitard is portrayed in the potentiometric map in Figure 4.13. The map was created using data only from Victoria as no monitoring bores are screened in the Upper Tertiary Aquitard in the South Australian part of the basin. In addition, the absence of measurements from the Narrawaturk Marl means that the map pertains mainly to the Gellibrand Marl.

Despite the sparse data distribution in some areas, coastward groundwater flow is confirmed in the Upper Tertiary Aquitard for the Gambier, Tyerndarra and Port Campbell Sub-basins. This is significant because quantitative hydraulic gradients have not been mapped for this aquitard before. The specific head distribution is discussed in relation to the aquifers in Section 5.3.



head contours in intervals of 50 mAHD; white lines are 10 mAHD contours; orange polygons are Upper Tertiary Aquitard outcrop; grey lines are surface water features; black arrows are groundwater flow directions; dashed red lines (short dash) are hydraulic groundwater divides; solid red lines are aquitard extent; red Figure 4.13. Potentiometric map for the Upper Tertiary Aquitard. Black crosses are locations of bores whose head measurements were used. Black lines are lines marked with a long dash are uncertain aquitard extent. This map was created using only Victorian data.

Also important is that a groundwater divide between the Corangamite region and the neighbouring Port Campbell and Barwon Downs Sub-basins is confirmed in Figure 4.13. These divides are significant because they mimic those of the watertable (Figure 4.4). In the Corangamite region the local flow system, contained mostly within the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers, directly overlies the Upper Tertiary Aquitard without any interaction with the Upper Tertiary Carbonate Aquifer, as the latter is absent in the area. The regional flow lines of the Upper Tertiary Aquifer are very similar to the local and intermediate flow systems of the watertable and this indicates the Upper Tertiary Aquitard's role in forming an internally drained basin in the Corangamite region. The low permeability of the aquitard (Table 4.1) prevents great vertical percolation and forces discharge in the low-lying centre at Lake Corangamite.

The flow within the Upper Tertiary Aquitard in the Barwon Downs Sub-basin also bears strong similarities to the shallow groundwater flow of the area. This is especially true in the south, where the watertable is contained within the outcropping Aquitard. Hydraulic head values higher than the watertable exist in the aquitard beneath Lake Colac, indicating an upwards gradient there. In combination with the localised watertable depression, Lake Colac is characterised as a discharge zone.

4.8 Flow in the Lower Tertiary Sandy Aquifer

The groundwater flow of the Lower Tertiary Sandy Aquifer is presented in the potentiometric map of Figure 4.14. This map incorporates some data from bores screened in sandy sections of the Mepunga Formation which occur near the Otway Ranges in the Port Campbell and Barwon Downs Sub-basins, despite the Mepunga Formation being part of the Upper Tertiary Aquitard in other regions. The map is the first to present hydraulic head data normalised for density for the entire aquifer system, including data from both Victoria and South Australia. Blake's map (1980) of the aquifer's potentiometric surface was not corrected for density and used only Victorian data. A map of head corrected for temperature for the aquifer was presented in Nahm's unpublished report (1985) on Victorian groundwater resources but this map was also only for the Victorian section. A modern map exists for the Victorian section (SKM, 2009a), but it is not density corrected. Love's maps (1993, 1994) were for the South Australian section only.

The direction of groundwater flow within the aquifer can be discerned from Figure 4.14 and, hence, the possible recharge and discharge zones are also apparent. However, because of the high relief in the structure contours of the aquifer (Figure 2.13) an analysis of the vertical flow components and their role in recharge and discharge processes is warranted. To aid with this discussion two more maps were created. Figure 4.15 shows the thickness of





sediment or rock overlying the top of the Lower Tertiary Sandy Aquifer: the overburden through which any recharging groundwater would have to flow to reach the aquifer, or discharging water to reach the ground surface. The vertical gradient $\left(\frac{dh}{dl}\right)$ between the Upper Tertiary Carbonate and the Lower Tertiary Sandy Aquifer was mapped using the head difference between the aquifers (dh) and the entire thickness of the Upper Tertiary Aquitard, including the thickness of the Clifton Formation (dl). Figure 4.16 shows the results with red tones indicating downward vertical gradients and blue indicating upward ones.

4.8.1 Barwon Downs and Port Campbell Sub-basins

These two sub-basins are discussed together because the groundwater flow paths within them originate in the same location. The division between them, shown in the maps, coincides with a groundwater mound located at the Barongarook High. To the north of the divide is the Barwon Downs Sub-basin; to the south, the Port Campbell Sub-basin in two lobes, dissected because of the distribution of the aquifer. The eastern lobe terminates near Gellibrand where the Lower Tertiary Sandy Aquifer is truncated at the point where the two basement highs are closest and shallowest. This structural feature is called the Gellibrand Saddle (Leonard *et al.*, 1983).

The main locations of recharge to the Lower Tertiary Sandy Aquifer for the Barwon Downs and the Port Campbell Sub-basins are the aquifer outcrops at the Barongarook High and on the northwestern flanks of the Otway Ranges (also observed by Blake, 1980 and Leonard *et al.*, 1983). Close to these areas the Lower Tertiary Sandy Aquifer is blanketed by the Upper Tertiary Aquitard and other surficial units which act to confine the aquifer in some regions and through which some recharge is also possible, depending on the vertical gradients and the thickness of the overburden. The latter is discernible from Figure 4.15 and the majority of that thickness near the outcrop is composed of the Upper Tertiary Aquitard. Unfortunately, the vertical gradients are not depicted in Figure 4.16 because the map only depicts gradients between the Lower Tertiary Sandy and Upper Tertiary Carbonate Aquifers and the latter is absent in the Barwon Downs region.

However, the vertical gradients are known in several parts of this region. In the Gerangamete and Murroon parishes, north of the groundwater divide between the two sub-basins, upwards gradients stronger than horizontal gradients are the normal situation when pumping from the Barwon Downs bore field is not in progress (Witebski *et al.*, 1995). In the eastern lobe of the Port Campbell Sub-basin is bore 108910, a flowing artesian bore screened in the Lower Tertiary Sandy Aquifer, confined by more than 100 m of Upper Tertiary Aquitard. As the watertable is below the surface at this location, the vertical gradient here is also upwards. These locations are very close, within approximately 10–15







Figure 4.16. Map of the vertical hydraulic gradient between the Upper Tertiary Carbonate Aquifer and the Lower Tertiary Sandy Aquifer, across the Upper Tertiary Aquitard. This map represents data from the Mepunga Formation in some places, where it consists of sandy lithofacies, but does not rely on data from the Clifton Formation of the Upper Tertiary Carbonate Aquifer. Red zones represent downward gradients and blue zones represent upward gradients.

km, of the divide where the gradient is downwards, demonstrating the shorter or more local nature of the flow systems in this region.

The upwards gradient and the constriction of flow at the Gellibrand Saddle forces some discharge to the Gellibrand River (Leonard *et al.*, 1983). However, it is not the final discharge zone for that flow system. The southwestward flow appears to be cut short at the boundary of the aquifer. However, as can be seen in Figure 2.1, the surface geology to the southwest of that flow line is dominated by the Upper Cretaceous Aquifer and flow within that unit is also southwestward (Figure 4.21). Therefore, it is postulated that the Upper Cretaceous Aquifer is in hydraulic connection with the Lower Tertiary Sandy Aquifer in the region surrounding Gellibrand and that this connection facilitates mixing of the previously confined groundwater in the north with more recently recharged water from the Upper Cretaceous Aquifer outcrop further south. This interconnection is feasible especially because of the complete absence or only very limited presence in the local area of Lower Tertiary Aquitard, which would normally act to confine the Upper Cretaceous Aquifer (refer to Figure 2.11). The interconnection in this area has been previously explicitly stated (Thompson, 1972; Blake, 1980; Leonard *et al.*, 1983), or implied by treating the Wangerrip Group as a single aquifer (Blake, 1980; refer to Table 2.1).

Following the radial flow from the primary recharge zone of the Barongarook High, the groundwater flow is generally towards the southwest in the Port Campbell Sub-basin and the northeast in the Barwon Downs Sub-basin. The extent to which the radial flow reaches westward marks the location of the groundwater divide with the Tyrendarra Subbasin. Once flow is disconnected from any easterly component, and is oriented NNE-SSW, the groundwater divide is established because flow lines are sub-paralell. The location of this groundwater divide is significantly eastward of the structural Warrnambool High between the sub-basins (c.f. Figure 2.2). However, it does coincide well with the maximum overburden (Figure 4.15).

The formation of regional flow systems becomes well developed directly from the Barongarook recharge zone as the aquifer dips deeper in both directions (Figure 4.15) and only local, rather than regional, discharge zones are present (e.g. those described above). The lack of regional discharge zones is confirmed by the persistence of downwards gradients almost to the coast in the western part of the Port Campbell Sub-basin in Figure 4.16. When the orientations of the patterns of the regional flow regime are compared to the morphology of the aquifer itself (Figure 2.13) the two are found to be very similar. This is not surprising, however, an important implication is the influence of faulting on the system. Normal faulting during the basin formation and subsequent reactivation and reversal have been the major mechanism through which the aquifer morphology has developed (c.f. Figure 2.21). Therefore, the occurrence and sense of movement on major faults in the region is one important control on groundwater flow.

Disrupting the regional flow system is a local- or intermediate-scale recharge zone in the southeast of the Port Campbell Sub-basin. This recharge zone is represented by a groundwater mound northeast of Port Campbell, associated with Lower Tertiary Sandy Aquifer outcrop and a small topographic high (Figures 4.14 and 1.2). As with local flow systems in other aquifers, this recharge zone creates some flow lines that are in the opposing direction of the regional flow system.

Further to this, recharge may be occurring via volcanic eruption conduits. It has been shown that the eruption centres are usually associated with fully-penetrating faults (e.g. O'Driscoll, 1960; Vanderzalm & Dillon, 2006) and/or the volcanic vents themselves may act as similar conduits if not completely plugged. This is especially likely to be the case with the phreatic or phreatomagmatic eruption centres, such as maars, which are underlain by diatremes (Singleton & Joyce, 1969; Ollier, 1988; e.g. Mrlina *et al.*, 2009). A considerable section of the Port Campbell Sub-basin contains many maars, including Lakes Purrumbete, Bullen Merri, Gnotuk & Elingamite (see Figure 7.8 for a map of the types of eruption centres). No groundwater mounds indicating recharge zones are visible in Figure 4.14 in the vicinity of these features. However, the paucity of closely-spaced data points precludes the detection of such features. North of the sub-basin, it has been suggested that the eruption centres are important for the recharge to the lower Tertiary Streatham deep lead aquifer (Raiber, 2008). Detailed investigations in the future may be able to determine whether eruption points are contributing significant recharge to the Lower Tertiary Sandy Aquifer in the Port Campbell Sub-basin.

Discharge mechanisms for the Lower Tertiary Sandy Aquifer, in all sub-basins, have been a problem for hydrogeologists studying the region for some time. Discharge is generally assumed to be offshore, but detailed investigations of the location and nature of discharge have not been performed in the past. The issue of discharge is not addressed here for the Port Campbell Sub-basin because the regional discharge in all regions is dealt with together in Section 4.8.4.

4.8.2 Tyrendarra Sub-basin

The groundwater of the Lower Tertiary Sandy Aquifer in the Tyrendarra Sub-basin flows south-southwest towards the coast from the basin margins (Figure 4.14). The aquifer is confined throughout the majority of the sub-basin (Figure 4.15), and it crops out only in a small zone in the valleys of creeks and rivers (e.g. Muddy Creek) in the northwest of the Sub-basin. The groundwater divides with neighbouring sub-basins are moveable boundaries, associated with subdued perturbations in the potentiometric surface and vaguely aligned with the structural features of the basin. The western and eastern ends of the sub-basin have sufficient numbers of observation points to confirm the groundwater flow paths. However, much of the central part of the Sub-basin is devoid of appropriate bore holes screened in the Lower Tertiary Sandy Aquifer except at a few locations along the coast. Therefore, the inferred hydraulic gradient across this region is not necessarily representative of the actual pressure distribution in the aquifer. Installation of observation bores in this region is crucial in furthering the understanding of the aquifer.

The recharge to the Lower Tertiary Sandy Aquifer in the western part of the Tyrendarra Sub-basin is around the eastern edges of the Dundas Plateau via areas of outcrop and through the relatively thin cover of other units (Figure 4.15; also asserted by Blake, 1980 and Leonard, 2003). Importantly, the Upper Tertiary Aquitard is absent in part of this area and the hydraulic gradient between the Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers is downward (Figure 4.16). SKM (2005) made estimates of recharge rates for the area and applied them to a calibrated numerical model. Their input value in the immediate vicinity of the outcrop along Muddy Creek was 150 mm/yr and the surrounding region of thin overburden was given a value of 50 mm/yr. Their study showed another estimated recharge zone of 150 mm/yr to the east of this, at the head waters of the Eumeralla and Moyne Rivers. However, that area is not considered to contain subcrop of the Lower Tertiary Sandy Aquifer by this study. However, the nearby volcanoes of Mt Napier and Mt Pierrepoint are within the aquifer limits and are probably conduits for recharge. Even eruption centres as far south as Mt Eccles are within the downwards gradient zone (Figure 4.16) and may be contributing to recharge along the regional flow line.

In contrast, the recharge mechanisms to the central and eastern parts of the Tyrendarra Sub-basin, including the region over the Warrnambool High, are more difficult to ascertain. In this region, the Lower Tertiary Sandy Aquifer is confined by about 200 m of Upper Tertiary Aquitard (Figure 2.14), but it is clear that recharge is possible, due to the persistence of downward hydraulic gradients in much of the north of the area (Figure 4.16). It has been demonstrated that interformational leakage is a viable mechanism for recharge to the Lower Tertiary Sandy Aquifer in the Warrnambool region by observations of leakage during pump tests (Shugg, 1976). However, given the considerable thickness of the Upper Tertiary Aquitard and its very low vertical hydraulic conductivity (Table 4.1), the flux is likely to be very low.

An alternative explanation, as discussed previously, is that the downward flux is in part accommodated by secondary porosity in the form of faults and volcanic pipes or conduits. There are a number of eruption centres in this part of the Sub-basin, the majority of them being maars (Figure 7.8). Further to this, although flow lines from the regional recharge zones to the east (Barongarook High) and west (Dundas Plateau) do not converge on the central region, the hydraulic gradients maintained by the influx in those regions is supporting the pressure regime in the middle of the Tyrendarra Sub-basin and over the Warrnambool High. That is to say, the water from the recharge zones does not ultimately reach the central region, but it does play a role in the groundwater flow of the area. This idea was touched on by Blake (1980) and Johns (1968) where they introduced the idea that the higher chlorinity or TDS of the groundwater indicates that the region is not effectively connected to the flushing of recharge water from the east. The reasons cited relate to fault-related thinning of the aquifer, possible lateral facies and related hydraulic conductivity changes and the obstacle to flow of structural highs (Johns, 1968; Blake, 1980). Their conclusions may imply that recharge from within the central part of the Sub-basin does not occur. This differs slightly from the idea that internal recharge rates are low. The issue of the groundwater quality in this area is discussed in Chapters 6 and 7.

The majority of the groundwater flow in the Lower Tertiary Sandy Aquifer in the Tyrendarra Sub-basin is part of a regional flow system. No major discharge features are present onshore, in fact, upwards hydraulic gradients are only observed in a small coastal portion of the Sub-basin (Figure 4.16). It is therefore assumed that groundwater flow continues offshore, despite the low hydraulic head values near Warrnambool. Offshore discharge is addressed further in Section 4.8.4.

4.8.3 Gambier Sub-basin

The flow of groundwater in the Lower Tertiary Sandy Aquifer within the Gambier Subbasin has been characterised by many authors already cited, most recently and comprehensively by Love *et al.* (1992b, 1993 & 1994). There are many similarities between the potentiometric map of this study and that of Love *et al.* (1993), including the mainly coincident positions of the zero head difference line (Love *et al.*, 1993) and the zero hydraulic gradient line (Figure 4.16). Given this similarity, most of the assertions made here about the flow system are also similar.

The main recharge zones in the Gambier Sub-basin for the Lower Tertiary Sandy Aquifer are: a large, gentle groundwater mound south of Penola; areas of aquifer outcrop and steep gradients in the river valleys of the Crawford, Stokes and Glenelg Rivers; and a smaller groundwater mound south of Strathdownie (refer to Figure 4.5 for locations). In addition, some recharge occurs in the north via the Kanawinka Fault (O'Driscoll, 1960; Kenley, 1962; Love *et al.*, 1993), although there is not a great deal of evidence for this in Figure 4.14 due to lack of data. These areas are well suited to effective infiltration of recharge to the confined aquifer because vertical gradients are strongly downward and overburden and Upper Tertiary Aquitard thickness are not extensive. Aquifer out crop is quite limited in this sub-basin but conditions are such that diffuse infiltration is prevalent, as discussed regarding the recharge for the Upper Tertiary Carbonate Aquifer in the same region. The thin overburden of the central recharge zone is related to the uplift of the Gambier Axis (see Chapter 2). The groundwater generally flows towards the coast from the recharge zones; although the flow emanating from the Gambier Axis is radial in nature. Flow at the northern margin is northwards into the Murray Basin.

The recharge zones along the rivers coincide with those identified by SKM (2005). Their estimates of leakage rates to the Lower Tertiary Sandy Aquifer, used as inputs for a calibrated flow model, were 150 mm/yr for the immediate vicinity surrounding the aquifer outcrop and 50 mm/yr for nearby area generally within the zone of less than 100 m overburden.

Harrington *et al.* (1999) analysed the groundwater of the Gambier Sub-basin using a compartmental mixing cell (CMC) model of solute transport in combination with a transient MODFLOW model of the flow. The model incorporated spatial and temporal change of physical and chemical attributes, and accounted for multi-layered flow and quantified vertical leakage to the Lower Tertiary Sandy Aquifer. The model produced results that agreed closely to measured values of hydraulic head and tracer concentrations and therefore the findings may be considered more reliable than those estimated from either physical or chemical methods alone. The recharge to the confined Lower Tertiary Sandy Aquifer, averaged across the entire zone of the Sub-basin upflow of the zero contour of the vertical hydraulic gradient map (similar to Figure 4.16), was found to be 2–9 mm/yr (Harrington *et al.*, 1999).

Recharge zones of a more local scale exist in the form of small highs in the potentiometric surface, one to the northwest of Mt Gambier and another south of Lake Leake. Both are in the red zone of the gradient map (Figure 4.16), and are in an area with very thin overburden (Figure 4.15), near a small exposure of the aquifer associated with the upthrown side of the Tartwaup Fault (Geological Survey of South Australia, 1951). However, the first high is juxtaposed to a small groundwater depression that resides in the upward gradient zone. This depression is associated with the crater of Blue Lake and indicates the process of baseflow to the Lake from the Lower Tertiary Sandy Aquifer. These perturbations in the potentiometric surface reflect local and intermediate flow systems and their position demonstrates how these systems interact with the regional flow system of the Lower Tertiary Sandy Aquifer.

Similar features in the potentiometric surface exist near the Glenelg River in areas where the aquifer crops out north of Dartmoor. This area is entirely within the regional downward gradient zone. However, the river does receive baseflow from the Lower Tertiary Sandy Aquifer via local flow systems, as indicated by some values of hydraulic head in the Lower Tertiary Sandy Aquifer that are higher than the watertable (c.f. Figure 4.4). The process is the same as that described for the baseflow contribution from the Upper Tertiary Carbonate Aquifer and the concept of the river receiving flow from both aquifers is supported by conclusions from an environmental tracer study (Allison *et al.*, 1975).

An interesting feature of the Gambier Sub-basin is the high absolute values of the vertical hydraulic gradients presented in Figure 4.16 when compared to values from the other regions. This is because the Upper Tertiary Aquitard in the Gambier Sub-basin has a relatively thin development (Figure 2.14), meaning that the *dl* component of the calculated gradient, $\frac{dh}{dl}$, is never as large as in other sub-basins. Hence, Figure 4.16 is only indicative of gradients relative to each other, as vertical gradients calculated from particular bores using individual screen intervals would differ significantly. Nonetheless, this relative relationship illustrates the ease with which recharge and discharge can occur to and from the Lower Tertiary Sandy Aquifer, comparatively speaking.

As with the other Sub-basins, the discharge of groundwater from the Lower Tertiary Sandy Aquifer in the Gambier Sub-basin is discussed in more detail below.

4.8.4 Regional Discharge from the Lower Tertiary Sandy Aquifer

As indicated above, discharge mechanisms for the Lower Tertiary Sandy Aquifer have always been a problematic research issue. Previous studies have confirmed the hydraulic connection with the ocean in the southward flowing sub-basins of the Otway Basin (Blake, 1980; Love *et al.*, 1992a; Love *et al.*, 1994; Harrington *et al.*, 1999) and have suggested that submarine discharge is the most likely form of regional discharge as no discharge zones are observed onshore. Similarly, a recent study of the Barwon Downs flow system has provided suggestions as to the regional discharge zone (near Bambra), rather than definitive answers (Petrides & Cartwright, 2006).

In addressing the issue of discharge for this aquifer, the change in boundary conditions due to sea level change, climate change or groundwater extraction must be addressed. Modelling has shown that such changes over the last 30,000 years are relevant (Harrington *et al.*, 1999). Blake (1980) postulated that the offshore discharge could either occur via faults or via seepage into the overlying formations, indicating that he believed the latter theory to be more plausible. These ideas are explored, along with the possibility that deep-cutting marine canyons on the continental slope expose the aquifer at the ocean floor and may allow seepage.

Submarine discharge from confined aquifers has generally been characterised as occuring at the intersection of the sea bed and the top of the aquifer, where the freshwatersaltwater interface propagates into the submarine extent of the aquifer (Figure 4.10; Taniguchi *et al.*, 2002). This particular part of the seabed can be in a submarine canyon (Essaid, 1990). However, if the tip of the interface has migrated into the aquifer, discharge may also occur through leakage to overlying layers (Essaid, 1990). The driver of the discharge is mixed convection: a combination of density-driven flow (free convection) and hydraulically-driven flow (forced convection) (Smith, 2004). The flow at the interface is of a cyclic nature wherein some of the water expelled as discharge from the submarine aquifer is reticulated seawater that had previously infiltrated the host rock (Smith, 2004; Taniguchi *et al.*, 2002).

A useful analogue for the Lower Tertiary Sandy Aquifer is the thick offshore sedimentary sequence of the Atlantic coastal plain. These strata are Cretaceous to Holocene in age and range from marine limestone to fluvial sand and silt (Trapp & Meisler, 1992). The presence of fresh or brackish groundwater in the sediments offshore has been confirmed by submarine sampling, groundwater with a chloride concentration less than 5,000 mg/L is found as far as 85 km offshore (Trapp & Meisler, 1992). The zone of fresher water is surrounded above by a thin mixing zone with seawater and bounded in the seaward direction by a thicker mixing zone (Meisler *et al.*, 1984). A similar situation has been observed at the north coast of Suriname, South America (Groen *et al.*, 2000).

One study applied the Ghyben-Herzberg Solution to the Atlantic coastal system and found that the measured interface position lies above the theoretical interface using predevelopment hydraulic head and below the theoretical interface according to post-development head values (Johnson, 1982). Thus, Johnson (1982) concluded that the interface is currently in a phase of landward/upward transition in response to recent groundwater extraction. However, subsequent numerical modelling, which has incorporated eustasy, has shown that the transience of the interface is more likely due to a lag in response to sea level rise throughout the Pleistocene (Meisler *et al.*, 1984; Trapp & Meisler, 1992). Modelling indicates that the interface is currently in equilibrium with a sea level approximately 15 to 30 m below current mean sea level (Trapp & Meisler, 1992). A very similar disequilibrium has been identified in the Soquel-Aptos Basin, California, through three-dimensional sharp interface modelling (Essaid, 1990).

Hence, these international analogues indicate that a lag in the response of the interface in a confined aquifer to changing boundary conditions can last for more than a millennium. Lags of such magnitude have been modelled in response to other types of boundary condition changes, such as groundwater abstraction (Bredehoeft & Durbin, 2009). Kooi *et al.* (2000) indicated that the rate of transgression and the geological setting are important factors affecting the time taken to achieve a new equilibrium, i.e. the time that a freshwater tongue can persist offshore. Specifically, the rapid global sea level rise of the last 10,000 years, i.e. during deglaciation of the last glacial maximum is conducive to a style of seawater intrusion where diffusive mixing from the seabed does occur but the lateral migration of the interface into the continental shelf is subject to a long lag (Kooi *et al.*, 2000). Also necessary for the situation observed in these basins is a wide continental shelf with a gentle slope, a thick aquifer with high hydraulic conductivity confined by a thick aquitard with low hydraulic conductivity, a geological framework quite common around the globe (Kooi *et al.*, 2000).

The Lower Tertiary Sandy Aquifer fits the above geological model well. The morphology of the continental shelf is visible in the digital terrain model (Figure 1.2), the Lower Tertiary Aquifer reaches a thickness of over 500 m offshore (Figure 2.12) and is consistently confined by a thick layer of marl (Upper Tertiary Aquitard; Figure 2.14). After the sea level high stand of the last interglacial, the Otway Basin region experienced oscillating but falling sea levels for about 100,000 years (Figure 3.10), culminating in a coastline more than 20 km offshore from its current position at 20,000 ka. It is therefore proposed in this study that, during this period, fresh groundwater existed up to or beyond the coastline in the Lower Tertiary Aquifer because the slow rate of regression allowed for the interface to adjust to a low sea level in a quasi-steady state manner (refer to Kooi *et al.* 2000). Conversely, after the last glacial maximum, the rate of transgression was much more rapid than the preceding regression and hence the interface has migrated inland, but not at a rate in equilibrium with sea level.

If it is accepted that the groundwater of the Lower Tertiary Sandy Aquifer discharges at the freshwater-saltwater interface, the position of the interface and the mechanisms or pathways of flow are of interest. Additionally, any physical evidence of the offshore discharge should be supportive of these theories.

The position of the interface could be well constrained by offshore pore water sampling in the Lower Tertiary Sandy Aquifer. However, inspection of offshore bore completion reports from the hydrocarbon industry revealed that such analyses are not attempted for the Lower Tertiary Sandy Aquifer because it is significantly shallower than the target reservoir formations (i.e. the lower portion of the Upper Cretaceous Aquifer). This may be an avenue for future research and industry cooperation.

The control of long-term sea level fluctuation on the interface has been discussed as far as possible without applying numerical modelling to the problem. Unfortunately, the analytical equations described in Section 4.6 (i.e. Equations 4.6, 4.7 and 4.8) are not directly applicable to confined aquifers. However, some of the relationships they define between the interface and other parameters are qualitatively upheld in confined situations. For example, two dimensional finite element modelling has confirmed that the interface moves landward and/or upward if the discharge flux is reduced (Volker, 1980). This is important for the Otway Basin case because of the presence of onshore groundwater


Figure 4.17. Diagram showing the parameters of Equation 4.12. The thick line is the interface that separates the fresh porewater (top) from the saline porewater (bottom), dashed lines are hydraulic head isopotentials in each fluid and bold arrows indicate the flow paths of the fluids.

abstraction (see Section 5.4.3), which shall eventually cause a decrease in discharge flux (Bredehoeft & Durbin, 2008). As this occurs, the imbalance in the system will be increased and, subject to a lag, will contribute to further landward migration of the interface in the future. The relationship between discharge and the interface is embedded in Hubbert's equation (1969) linking the angle of the interface to the hydraulic gradient in the aquifer:

$$\sin\gamma = \frac{1}{g} \left(\frac{\rho_s}{\rho_s - \rho_f} \cdot \frac{dh_s}{dL} - \frac{\rho_f}{\rho_s - \rho_f} \cdot \frac{dh_f}{dL} \right)$$
(4.12)

where γ is the angle above horizontal to the underside of the interface; $\frac{dh_s}{dl}$ and $\frac{dh_f}{dl}$ are the hydraulic gradients in the freshwater and saltwater parts of the aquifer respectively and where L is distance in the orientation of the interface (Figure 4.17).

If it is assumed that the salt water and the fresh water are both flowing under the influence of similar magnitude but opposite hydraulic gradients, the relationship between the absolute value of that gradient and the angle of γ is positive (Figure 4.18). The equation is such that, the weaker the hydraulic gradient $(\frac{dh}{dL})$, the smaller the angle of the interface (γ) and therefore the further into the aquifer the toe of the interface (x_T) can protrude.

Despite the importance of this understanding, it does not better constrain the length of the lag or the position of the interface. It is concluded above that the lag is mainly in response to long term sea level change, requiring a lag in the order of millennia. However, this idea is challenged by the idea that diurnal tidal fluctuations have been observed in



Figure 4.18. Plot of γ , the angle of the freshwater-saltwater interface (see Figure 4.17) with variations in hydraulic gradient (dh/dl). The gradients in the fresh and salt water were assumed to be equal and opposite. The data, derived from Equation 4.12, approximate a straight line as shown (R² provided).

the hydraulic head measurements of bores tapping the Lower Tertiary Sandy Aquifer at Portland (Shugg, 1976); meaning that lags are very short. Further research is needed to resolve this issue but it is possible that the fluctuations measured at Portland where not due to tidal loading but to barometric loading of the water column in the bore (Sara, 2003).

As to the position of the interface, one marker of the discharge zone is the presence of upward hydraulic gradients. The vertical gradients mapped between the Upper Tertiary Carbonate Aquifer and the Lower Tertiary Sandy Aquifer Figure 4.16) reveal the position of the hinge line of the flow system, i.e. where the vertical gradient is close to zero and flow is neither downward or upward but is horizontal. The distance up gradient of the hinge line to the regional recharge zone should be similar to the distance down gradient of the line to the regional discharge zone. This concept is illustrated in an idealised cross section in Figure 4.19. Judging by the position of the hingeline in Figure 4.16, the discharge zone may lie approximately 50 km offshore in Port Campbell or Tyrendarra Sub-basin, or about 20–30 km in the Gambier Sub-basin (c.f. generalised cross section in Figure 4.19). These estimates are consistent with the length of the modelled zone down gradient of the hingeline in the study of Harrington *et al.* (1999).

Four main types of discharge mechanisms for the Lower Tertiary Aquifer are depicted in Figure 4.19; they are: diffuse leakage to overlying units, leakage via faults or volcanic pipes, and direct discharge to the ocean from the exposed aquifer surface at the sea floor. As they are represented in this schematic diagram, each mechanism corresponds to a



Figure 4.19. Schematic and simplified cross section showing recharge and submarine discharge pathways for the Lower Tertiary Sandy Aquifer in their current possible locations: \mathbf{a} -via volcanic pipes; \mathbf{b} -via faults; \mathbf{c} -diffuse leakage; and \mathbf{d} -direct leakage in submarine canyons. Discharge pathways are associated with differing potential interface locations (red lines); each line depicts a sharp interface but represents a zone of mixing between saline and fresh water in the aquifer. The transgression since the last glacial maximum is discernible in the difference between the two sea levels depicted; the hinge zone and discharge areas would have been further offshore at 20 ka and are currently still moving landward. Horizontal scale is approximate; vertical exaggeration is about 1:20. Geology was adapted from Figure 2.6 (this study) and Figure 2.19 (Moore *et al.*, 2000).

freshwater-saltwater interface in a different possible location; in reality, only one diffuse interface position is valid and a combination of mechanisms is likely.

The expedition of discharge in offshore areas probably occurs via faults or volcanic pipes. Unfortunately, lack of information regarding offshore eruption centres (Refer to Figure 2.16 for offshore volcanic outcrop) and the composition of volcanic pipe infill means that the viability of the pipes as pathways is still largely unknown. Leakage along faults is inhibited by the plasticity of the marl of the Upper Tertiary Aquitard (Blake, 1980; Petrides & Cartwright, 2006) and the termination of many faults below or within the aquitard, rather than full penetration (Figure 4.19; Finlayson *et al.*, 1996; Moore *et al.*, 2000; Palmovski *et al.*, 2004). However, several pieces of evidence indicate that faults are an effective discharge pathway. First, it is known that some onshore faults are associated discharge springs (e.g. Tartwaup Fault; Sprigg, 1952; O'Driscoll, 1960). Faults and joints are particularly important pathways of leakage between the Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers where the Upper Tertiary Aquitard is thin, i.e. in the Gambier Sub-basin (pers. comm. J. Lawson, 2006).

Second, evidence exists for leakage of natural gas from traps below the Lower Tertiary Aquitard. This evidence includes gas inclusions in quartz grains of spent reservoir intervals (pers. comm. J. Underschultz, 2009), gas chimneys interpreted in seismic sections (Boult *et al.*, 2006), and observations of current submarine seeps (Appendix K contains video footage of a leak offshore from the Otway Ranges). Remaining commercial gas reservoirs have been filled by migration along faults and demonstrate that not all faults in the basin are leaky (Mehin & Link, 1994). Generally, the longer faults, or those that are close to fully-penetrating, are the ones that were reactivated during the basin's inversion periods (Chapter 2).

Third, evidence for these long, penetrating faults exists offshore. Seismic lines are direct evidence of fault morphology (Figures 2.17, 2.18, 2.19 and 2.20). Although these interpretations do not show many faults that penetrate the Upper Tertiary, high resolution bathymetry (restricted-access 250 m grid raster; provided by Geoscience Australia) clearly shows the expression of faults at the top of the continental slope, near the heads of the submarine canyons. Unfortunately, these features are not visible in Figure 1.2 (1000 m resolution). The features are deduced to be expressions of faults rather than geological contacts because they cross cut elevation contours.

Another very plausible discharge mechanism for the Lower Tertiary Aquifer is diffuse leakage through the overlying strata (Blake, 1980; Love *et al.*, 1993; 1994). Relatively less retardation of diffuse leakage occurs in the Gambier Sub-basin, where the Upper Tertiary Aquitard has a thin development (Figure 2.14). Upwards flow from the aquifer into other units can be induced by density gradients between the groundwater and seawater that has permeated the aquifer matrix. Once groundwater flows into an adjoining aquifer or aquitard, its fate is determined by the flow mechanisms within the new domain.

The last option for discharge is direct flow of groundwater from the Lower Tertiary Sandy Aquifer into the ocean. The prime location for this occurrence is where the aquifer is exposed in the submarine canyons. In some cases, the aquifer pinches out before reaching the edge of the continental shelf (Figure 2.7), precluding this mode of discharge. Additionally, discharge can only occur in the canyons if the freshwater-saltwater interface is positioned very close to the sea floor. This may be the case where the shelf is narrow and hence the distance from the hingeline to the continental slope is short (Figure 4.19). The canyon fill of the Otway Basin submarine canyons is likely to be rich in sand (Leach & Wallace, 2001) and therefore has a hydraulic conductivity conducive to allowing seepage.

Canyon seepage is supported by the theory of Boult *et al.* (2006), who propose several pieces of evidence that the modern canyons form a preferential discharge pathway for hydrocarbons, including gas but also asphaltite that eventually forms tar mats that become stranded on beaches of the Gambier Sub-basin. Faults are expected to play an important role as a migration pathway prior to the hydrocarbons reaching the canyon incision (Boult *et al.*, 2006).

It is the conclusion of this thesis that all the above mechanisms are viable for the submarine discharge from this aquifer. In seeking support for this conclusion, oceanographic data from the offshore Otway Basin region was analysed for anomalies that may represent inputs groundwater in the ocean water column. The data used were conductivity temperature depth (CTD) profiles of the ocean water measured by *in situ* probes at numerous locations by CSIRO research cruises distributed over the years from 1948 to 2001. The data were sourced from MarLIN: the CSIRO Marine and Atmospheric Research Laboratories Information Network (http://www.marine.csiro.au/marlin/).

The signals searched for in the CTD data were anomalously high temperature and/or anomalously low salinity because these are the characteristics that differentiate groundwater of the Lower Tertiary Sandy Aquifer from seawater (refer to Chapter 6 for more detail on the salinity of the groundwater). However, these characteristics are likely to be muted somewhat by the mixing with seawater that occurs as part of two of the above discharge processes (Figure 4.19), making anomalies harder to identify. Reported mixing ratios between groundwater and recycled seawater in discharge include volume ratios of 2:8 and 4:96 (Smith, 2004). Nonetheless, depressions in bathymetry thought to be submarine groundwater springs offshore from Townsville, Queensland, have been found to contain detectable amounts of freshwater (Stieglitz, 2005). Therefore, fresh groundwater discharge can be detected in the submarine environment if conditions are favourable.

If discharge is from a point source, the discharge water, containing some component of

groundwater, rises in a plume due to its buoyancy (low density). The plume rises to a level where its density is equal to surrounding water and then begins to spread laterally (Speer & Helfrich, 1995). This idea is supported for the Otway Basin by the presence of low salinity anomalies in the upper 2 m of the ocean water column offshore from the Gambier Sub-basin, thought to be groundwater seeps derived from the ocean floor (Wylie, 1969; these anomalies are close to the coast and hence are likely to represent discharge from the Upper Tertiary Carbonate Aquifer, rather than the Lower Tertiary Sandy Aquifer). During this density-driven flow, buoyancy of the plume water is reduced because of a decrease in temperature and increase in salinity. Temperature is decreased rapidly through conduction and convection; whereas salinity can only be increased by advection and diffusion, both of which are inhibited by the relatively rapid buoyant flow of the plume. In accordance with these theoretical points, only salinity anomalies indicating groundwater were found in the CTD data.

Significant low-salinity anomalies where found in the CTD data from most cruises, at variable depths and at locations on the shelf and slope. Additionally, the in situ density was calculated from the CTD data for each profile (using the algorithm of Fofonoff & Millard, 1983) and the results showed that these anomalies were usually associated with density inversions, indicating instability of the water column. Therefore, although no temperature anomaly was observed, the readings are thought to represent diffuse plumes of groundwater discharge rising through the water column and slowly mixing with the seawater. The best example of the anomalies found comes from two profiles (locations 97 and 98) of the 1991 cruise FR1991_04 (Figure 4.20). The localities are approximately 15 km apart and are situated at the toe of the shelf, near the head of submarine canyons, about 90 km south of Warrnambool. The measurements taken at locality 98 show the anomaly present approximately 700 m above the depth at which it was encountered at locality 97. Considering the profiles are so close and that they were conducted within 3 hours of each other, it is possible that the shape and depth of the anomalies at each location track the rise and diminishment of the discharge plume (Figure 4.20). It has been noted by other authors that lateral transport of the plume as it emerges from the ocean floor is subject to marine currents (Wylie 1969; Boult et al., 2006; pers. comm. J. Middleton, 2006).

Very similar density inversions and salinity anomalies have been found in the ocean water of the Great Australian Bight (McClatchie *et al.*, 2006; 500 km north west of the offshore Gambier Sub-basin). The anomalies were found in a variety of settings, generally more than 50 km from the coast and at a depth range of 40–60 m, close to the ocean floor. McClatchie *et al.* (2006) deduced that the density inversion was sustained by diffuse leakage of groundwater from the ocean floor.



Figure 4.20. CTD profile showing low salinity anomalies in ocean water that correspond to density inversions offshore from Warrnambool, Victoria. Pressure is used as a proxy for depth; the values (decibars and metres) are roughly equivalent, although the relationship changes with increasing pressure.

Overall, theoretical and practical observations in the Otway Basin and in similar offshore settings around the world corroborate to provide convincing evidence that the groundwater of the Lower Tertiary Sandy Aquifer flows some distance beyond the coastline and discharges to the ocean via diffuse leakage and flow though faults and other conduits.

4.8.5 Groundwater velocity and residency in the Lower Tertiary Sandy Aquifer

Hydraulic estimates of groundwater velocity and residency in the Lower Tertiary Sandy Aquifer were made for each of the major sub-basins following the same methods outlined in Section 4.6.1 for the Upper Tertiary Carbonate Aquifer. This approach uses Darcy's Law and therefore forces the assumption of steady state conditions.

The Lower Tertiary Sandy Aquifer increases in thickness from the marginal recharge zones towards the current coastline and then decreases thickness offshore (Figure 2.12). The result of steady state flow through such an aquifer system is that groundwater velocity must change to accommodate the same flux at the beginning, middle and end of the flow paths. Hence, the velocity in the shallow, thin recharge zone of the aquifer is possibly faster than that at the hinge-line in the deep part of the aquifer (e.g. Lawrence, 1975). Conversely, most aquifers are anisotropic, therefore, the vertical components of flow at the recharge zone, which are largely absent in the hinge zone, are subject to the vertical hydraulic conductivity. In sedimentary aquifers, this is likely to be lower than the horizontal hydraulic conductivity (Heath, 1983).

For these reasons, the groundwater velocity was estimated from horizontal gradients near the hinge line using horizontal hydraulic conductivity (refer to Section 4.6) and the velocity thus calculated may not be representative for all sections of the groundwater flow path. The results of the analysis for the Lower Tertiary Sandy Aquifer are presented in Table 4.6.

The results show that similar ranges of groundwater velocity are estimated for the Lower Tertiary Sandy Aquifer in the Tyrendarra and Port Campbell Sub-basins, and that the velocity in the Gambier Sub-basin is somewhat slower. These hydraulic velocities were used to calculate the residence of groundwater that has flowed from the recharge zone to a point at the coast on each flow line, namely: Portland, Port Macdonnell and Port Campbell. Thus, the hydraulic residence times presented in Table 4.6 are for groundwater travel along the entire onshore flow path, despite the gradient used in the calculation being only derived from the hinge zone. The values are compared to corrected groundwater radiocarbon ages from the same locations.

In the case of each sub-basin, the hydraulic residence times for the flow paths of given

ın Figure 4.5														
Sub-basin]	Horizontal A h/dl at 1ingeline	l K (max)	K (min)	n _e (min)	n _e (max)	q (max)	q (min)	v (max)	v (min)	Location for comparison	Distance to location from recharge zone	Residence time at location (max)	Residence time at location (min)	¹⁴ C Age (corrected)
		[m/day]	[m/day]			[m/day]	[m/day]	[m/day]	[m/day]		[km]	[years]	[years]	[¹⁴ C years]
Tyrendarra	0.0002	20	13	0.15	0.25	0.004	0.003	0.03	0.01	Portland	75	19,930	7,773	24,800
Tyrendarra	0.0004	20	13	0.15	0.25	0.009	0.006	0.06	0.02	Portland	75	8,785	3,426	24,800
Tyrendarra	0.0017	20	13	0.15	0.25	0.033	0.022	0.22	0.09	Portland	75	2,384	930	24,800
Gambier	0.0009	8	2	0.15	0.25	0.007	0.002	0.05	0.01	Pt Macdonne	1 75	27,653	4,148	> 30,000
Gambier	0.0012	8	2	0.15	0.25	0.010	0.002	0.07	0.01	Pt Macdonne	11 75	20,870	3,131	> 30,000
Pt Campbell	0.0004	11	လ	0.15	0.25	0.004	0.001	0.03	0.005	Pt Campbell	20	11,777	1,927	14,200
Pt Campbell	0.0013	11	က	0.15	0.25	0.015	0.004	0.10	0.02	Pt Campbell	20	3,408	558	14,200
Pt Campbell	0.0027	11	က	0.15	0.25	0.030	0.008	0.20	0.03	Pt Campbell	20	1,662	272	14,200

Table 4.6.	. Hydraulic groundwater velocity and residence times in the Lower Tertiary Sandy Aquifer. dh/dl—hydraulic gradient; K—horizontal hydraulic
conductivity	y (standardised to 20° C); n _e -effective porosity; q-Darcy flux (Equation 4.10); v-hydraulic groundwater velocity (Equation 4.11). Distances were
estimated fro	rom Figure 4.14. Radiocarbon ages are presented in Chapter 7, the age for Port Macdonnell is from Love et al., (1994). Locations are shown on a map
in Figure 4.5	5.

length are shorter than the corrected radiometric residence times. There are two possible reasons behind this difference. First, part of the discrepancy may be due to diffusion of older water from surrounding aquitards into aquifers having a marked effect on radiometric ages, which cannot be accounted for in hydraulic analyses (Bethke & Johnson, 2002; 2008). This phenomenon is often responsible for radiometric age in aquifers being greater than expected from hydraulic velocity estimates (Bethke & Johnson, 2002).

Second, in calculating the hydraulic residence times for this study, the hydraulic groundwater velocity at the hinge line has been applied to the entire flow path. As mentioned above, this is not an entirely valid assumption. If the radiometric ages are assumed to be representative of the true groundwater age, then the hydraulic residence time is too short. This would indicate that the hydraulic velocity is faster than the average actual groundwater velocity over the flow path. Therefore, the vertical infiltration to the aquifer at the recharge zone is at a velocity slower than that measured for the horizontal flow at the hinge line. Notwithstanding the discrepancy, the hydraulically-derived groundwater velocities are a good indication of the variability of velocity present in the aquifer around the hinge zone.

A discrepancy was also recorded between radiometric and hydraulic residence times for the Lower Tertiary Sandy Aquifer in the northern Gambier Sub-basin by Love *et al.* (1992a, 1994). However, the relationship between the ages observed by Love *et al.* (radiometric < hydraulic; 1994) is the opposite to that shown by the calculations in this study (Table 4.6; radiometric > hydraulic). This is because the calculations of each study are slightly different and, accordingly, the discrepancies are attributed to different factors. In the study of Love *et al.*, the calculated ¹⁴C residence time for groundwater for piston flow from the hinge line to the coast (c. 50 km) is 12,800 years; whereas the hydraulic residency over same distance is 49,000 years (1994). The older hydraulic ages were attributed to the current hydraulic gradient being low due to a sea level high stand; the younger radiometric ages were attributed to a chemical signature remaining from faster velocities of the past that were due to a higher hydraulic gradient of the low sea level phase of the last glacial maximum (Love *et al.*, 1994).

As with the Darcy's Law flow analysis of the Upper Tertiary Carbonate Aquifer, the groundwater velocities derived here provide a timeframe relevant to the evolution of the groundwater in the Lower Tertiary Sandy Aquifer as a consequence of palaeoclimate fluctuations and water-rock interaction.

4.9 Flow in the Upper Cretaceous Aquifer

The groundwater flow system of the Upper Cretaceous Aquifer is depicted in Figure 4.21. No suitable bores were available for the Gambier Sub-basin and few are present in the Tyrendarra Sub-basin and the western part of the Port Campbell Sub-basin. However, this is the first potentiometric map of the Upper Cretaceous Aquifer to be produced independently of data from the Lower Tertiary Sandy Aquifer and is the first to represent all the density corrected data available. It depicts data from the upper part of the aquifer only, namely: Pebble Point Formation, Moomowroong Sand Member, Timboon Sand and Paaratte Formation.

Despite the paucity of data, some generalised flow paths can be inferred for the western parts of the aquifer which agree well with what is known about the flow in the overlying units. For these regions flow is probably coastward and recharge presumably occurs through the overlying strata, as there is no outcrop. It is hard to judge at what points in the flow system the vertical hydraulic gradients are downward or upward. However, it is reasonable to assume that recharge and discharge mechanisms for the Upper Cretaceous Aquifer are similar for those of Lower Tertiary Sandy Aquifer.

The number of bores from which the potentiometric surface can be mapped is significantly greater in the east of the basin. This is mainly because the Upper Cretaceous Aquifer lies close to the surface as a result of the inversion which created the Otway Ranges. For the same reason a significant exposure of the aquifer exists in the Port Campbell Subbasin. The aquifer becomes progressively more deeply buried to the south and west of the Otway Ranges in the Port Campbell Sub-basin and to the northeast of the Barongarook High in the Barwon Downs Sub-basin.

The main recharge zone for both regions is the Barongarook High, from which the groundwater flow emanates in three main directions: semi-radially to the north; to the east and then south over the Gellibrand Saddle; and to the west and then south. The two latter flow systems reunite in the area where the aquifer is exposed, to the south of the Otway Group basement outcrop at the Barongarook High. In this region the flow mixes with water recharged directly into the aquifer outcrop, some of which is sustained by westward runoff from the Otway Ranges.

The position of the groundwater divide between the two Sub-basins is not definitive but it is significant in that it highlights the south bound flow of groundwater over the Gellibrand Saddle, towards the outcrop area. This flow system has the same orientation as that of the Lower Tertiary Sandy Aquifer and this is the area where they are hydraulically connected due to the absence of the Lower Tertiary Aquitard. Further along the flow system, to the south and west, the two aquifers are separated by the Lower Tertiary



Figure 4.21. Potentiometric map for the Upper Cretaceous Aquifer. This map depicts hydraulic head from bores (marked by crosses) screened in the upper part of the aquifer; refer to Table 2.1. Black lines are head contours in intervals of 50 mAHD; white lines are 10 mAHD contours; blue polygons are Upper Cretaceous Aquifer outcrop; grey lines are surface water features; black arrows are groundwater flow directions; dashed red lines (short dash) are hydraulic groundwater divides; solid red lines are aquifer extent; red lines marked with a long dash are uncertain aquifer extent. The basement outcrop of the Barongarook High is shown as transparent white polygons. Aquitard and their flow is accordingly refracted. As with the other areas, discharge is assumed to occur offshore.

4.10 Synthesis

This chapter has provided a complete overview of the physical hydrogeology of the Otway Basin at different scales and in as many hydrostratigraphic units as the data allow. Five potentiometric maps, produced from density corrected hydraulic head data, are used to aid the description of the flow systems of the basin; in some instances, the systems are mapped or described for the first time. Regional flow systems are generally recharged inland near basement outcrop and flow towards the ocean; regional discharge is submarine or coastal. Local flow systems are more complex and involve flow paths that can be opposite to regional coastward flow. Complimenting the maps, calculations are made of recharge, groundwater velocity and the position of the freshwater-saltwater interface. The work presented in this chapter supports several conclusions relating to the geological, hydraulic, climatological and eustatic controls on groundwater flow in the basin.

Geological controls

The recharge zones and groundwater divides of the watertable are strongly associated with the eruption centres of the Pliocene-Quaternary Volcanic Aquifer because the centres are dominated by lithofacies such as scoria and tuff that have high hydraulic conductivity. Accordingly, variation in recharge magnitude with lithology or related soil type is observed through watertable fluctuation calculations. Although not previously suggested for the Otway Basin, volcanic pipes, especially diatremes underlying maars, are thought to be recharge and discharge conduits for the deep confined aquifers, e.g. Lower Tertiary Sandy Aquifer.

Karst environments of the Plio-Pleistocene Heterogeneous Aquifer and the Upper Tertiary Carbonate Aquifer provide conduits for rapid vertical infiltration and also for discharge at major coastal springs. The effect of karstification on topography has resulted in the development of complex localised flow patterns in the shallow levels which interact with more uniform intermediate-scale flow systems at depth.

Faults act as discharge pathways onshore and offshore for both unconfined and confined aquifers, although not all faults are plausible pathways. In addition, faults are responsible for the current aquifer structure, which in turn has an influence on flow direction, especially in the Port Campbell and Barwon Downs Sub-basins.

The point source recharge and discharge described above are accompanied by diffuse infiltration and leakage and which is controlled by the thickness and hydraulic conductivity of the aquitards, most importantly the Upper Tertiary Aquitard. The Gambier Sub-basin has the thinnest extent of this aquitard and hence diffuse recharge and discharge can occur to and from the Lower Tertiary Sandy Aquifer with relative ease.

Hydraulic controls

The hydraulic controls on groundwater flow have been quantified through the creation of potentiometric maps and calculations of groundwater velocity. In the shallow or surficial aquifers, hydraulic head levels are often close to 0 mAHD near the coast. This demonstrates the hydraulic connection with the ocean and indicates that current discharge zones are subject to the position of the freshwater-saltwater interface. The moderate hydraulic gradient near the coast, especially in the Gambier Sub-basin, creates susceptibility to seawater intrusion. Areas of high diffusivity (often associated with karstification) have been affected already and further landward migration of the freshwater-saltwater interface is expected because of an hydraulic lag.

Calculations using Darcy's Law indicate that the vertical infiltration to the Lower Tertiary Sandy Aquifer is slower than the groundwater velocity at the hinge zone. Generally, the flow velocity for this aquifer is similar in the Tyrendarra and Port Campbell Sub-basins and tends to be slower in the Gambier Sub-basin. This slower vertical flow counteracts the faster recharge and discharge flow noted earlier for the Gambier Sub-basin.

Climatological controls

Fluctuations of hydrographs for unconfined aquifers indicate that winter precipitation contributes seasonally to recharge and no major temporal change in the magnitude of seasonal recharge is discernible from the data. However, a decline in hydraulic head of shallow aquifers indicates that storage is being reduced. It is concluded that this loss of groundwater is partly due to an increase in the volume of water discharged via evapotranspiration, notably during the summer months. These observations fit well with the declining P/E ratio recorded for approximately the last 200 years (Chapter 3).

Eustatic controls

Modern and future sea level rise (Chapter 3) could be detrimental to shallow groundwater resources as it is likely to cause further landward encroachment of the freshwater-saltwater interface in the unconfined aquifers. Seawater intrusion has been identified by the Border Groundwaters Agreement Review Committee (2008) as a serious risk for irrigators and groundwater dependent ecosystems (e.g. Piccaninnie Ponds).

For the confined Lower Tertiary Sandy Aquifer, eustatic sea level rise since the last

glacial maximum (20 ka; Chapter 3) is still relevant to the position of the freshwatersaltwater interface. It is the conclusion of this study that fresh groundwater exists some 20–50 km offshore within the aquifer; accordingly, the lag in response to eustasy is in the order of millennia. This conclusion is based in part on the similarity of the Otway Basin hydrogeology to situations in other basins around the world. The freshwater-saltwater interface is therefore not in equilibrium and will continue to migrate inland into the future.

Thus, this chapter provides a physical groundwater flow framework, within which the hydrogeochemical evolution of the groundwater can be assessed in Chapters 6 and 7. In addition to the conclusions described above, further investigation of climatological changes and hydraulic connection between hydrostratigraphic units is made in Chapter 5, along with exploration of the effects of land use change, hydrocarbon development and groundwater extraction.

Chapter 5

Interaquifer flow and aquifer stress

5.1 Introduction

In this chapter the understanding of the flow system gained from Chapter 4 is further developed. First, the regional-scale interaquifer interaction is addressed, aided by inferred flow lines along cross sections. The cross sections used are those first presented in Chapter 2 to display the relationships between the hydrostratigraphic units in each of the major sub-basins. Second, the aquifers' response to stresses is explored by way of a systematic analysis of hydrograph trends. Linear trends of hydrographs are presented in maps for several hydrostratigraphic units. The rising and falling trends are analysed with specific reference to regional and local causal factors, including groundwater extraction, long term climate change, sea level fluctuations and land use change.

5.2 Methods

5.2.1 Cross section production

In Section 4.2, the methodology of the creation of the watertable and potentiometric maps was provided. The cross sections provided in this chapter were created by using these maps and hence, the data handling methodology is not repeated. Hydraulic head values from the maps were transcribed to the cross sections by hand as points. To locate the points on the cross sections, the lateral position of a point was defined as the intersection of a head contour from a map with the line of the section. The depth location of a point was determined by the depth on the cross section of the hydrostratigraphic unit represented in the map. The resultant net of hydraulic head points in the cross sections where contoured by hand, thus creating isopotentials. Groundwater flow lines orthogonal to the isopotentials provide an indication of the inter-aquifer flow regime.

This method is quantitative only in an indicative sense, as the exact depth of screen intervals was not used for transcribing the head measurements from the maps to the sections. Additionally, the cross sections are vertically exaggerated and so the vertical and horizontal components of the hydraulic gradients and groundwater flow paths are not presented precisely.

5.2.2 Hydrograph analysis

The historical records of water levels for 598 bores thoughout the Victorian section of the Otway Basin, comprising 54,027 measurements, were analysed for rising or falling trends. Measurements for each bore were plotted as hydrographs and were visually inspected for anomalous readings. Bore drawdown recorded in hydrographs was noted but was considered inappropriate for calculation of long term trends in head. Once erroneous or anomalous data had been discarded, the measurements for each borehole were used to derive equations of lines of best-fit using *Microsoft Office Excel*. The slopes of these lines express the long term rate of change in head over time; trends are either rising (positive slope), neutral (slope close to zero) or falling (negative slope). Some representative hydrographs previously presented for the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers (Figure 4.6, Section 4.4) are used as an indication of the type of data from which the linear slopes were calculated, as well as some typical hydrographs for the Upper Tertiary Carbonate, Lower Tertiary Sandy and Upper Cretaceous Aquifers (Figure 5.1).

Steps were taken to eliminate short term trends from the dataset. Measurements clearly affected by bore drawdown related to known groundwater extraction were not included in the trend calculations, and, in order to eliminate the possibility of seasonal fluctuations influencing the trends, records for bores that were less than 2 years in duration were discarded. Lastly, records for a bore that consisted of fewer than 5 measurements were also discarded.

No records available to this study for any one bore were longer than 36 years; the monitoring records used span the years 1974–2006. There are hydraulic head levels recorded for the region prior to 1974, however, they are not appropriate for this analysis. Such readings may not be recorded digitally in the Victorian Groundwater Management System or they may consist of a single reading in a bore at the time of bore completion which cannot be extrapolated into a trend, as no subsequent readings were made for that bore. These shortcomings of the dataset indicate that, for future work to be more rigorous, there is a need to digitise pre-1970s records and possibly to spatially integrate "loose" data points.



Figure 5.1. Typical hydrographs for the Upper Tertiary Carbonate Aquifer (turquoise), the Lower Tertiary Sandy Aquifer (blue) and the Upper Cretaceous Aquifer (dark blue) and their linear trendlines. Listed above each hydrograph are: the bore location (parish); bore number; and rate of change in head (m/yr). The head axis has been truncated to fit all the data in one graph, but the scale remains consistent for each hydrograph. Note that the trend for the Murroon bore is independent of bore drawdown and those of the Belfast and Pirron Yaloak bores are independent of the spurious data points.

5.3 Interaquifer flow

Having discussed each aquifer's flow regime in detail in Chapter 4, it is now prudent to provide a more general and integrated analysis of the interaquifer flow patterns. The insight into these patterns is provided by three flow cross sections, one in each of the Gambier, Tyrendarra and Port Campbell Sub-basins. The cross sections are presented below in Figures 5.2, 5.3 and 5.4; their locations are the same as the geological cross sections presented in Chapter 2 and are marked in Figure 2.3.

5.3.1 Gambier Sub-basin

The hydrogeological cross section in the Gambier Sub-basin shows the interaction between local intermediate and regional flow lines. There are at least four distinct local recharge zones interspersed with local discharge zones associated with the Glenelg River and a tributary (Figure 4.5). These local flow systems are contained within the Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers and illustrate the hydraulic connection between these Aquifers that has been noted by several other authors in the past (e.g. O'Driscoll, 1960; Waterhouse, 1977; Love *et al.*, 1993). As the Upper Tertiary Carbonate Aquifer thickens towards the coast, flow within it becomes more regional in nature, highlighting the importance of the recharge zone that lies halfway along the section. The isopotential contours within the Upper Tertiary Carbonate Aquifer have different slopes according to their position within the regional flow system. Near the coastline the contours are roughly vertical, indicating ongoing horizontal flow and eventual submarine discharge. However, some shallow, more local flow lines also exist near the coast allowing discharge there.

Recharge to the Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer occurs via two main zones. The first is in the highlands at the edge of the Dundas Plateau, at the right hand margin of the cross section. The second is associated with the small pressure high in the Lower Tertiary Sandy Aquifer south of Strathdownie (Figure 4.5), approximately 25 km along the cross section. This mound forces the formation of an intermediate flow system which involves some flow against the regional flow direction. This is shown in the cross section where it can be seen that some flow in the Lower Tertiary Sandy Aquifer is northward and some is into (denoted as \otimes) and out of (denoted as \odot) the page (Figure 5.2). This recharge zone is near the location of several nested borehole sites that have recorded downwards hydraulic gradients within the Lower Tertiary Sandy Aquifer (Blake, 1980).

Offshore, the Upper Tertiary Aquitard thins and an upward gradient enables leakage from the Lower Tertiary Sandy Aquifer to the Upper Tertiary Carbonate Aquifer, which in





turn ultimately discharges to the ocean. Movement of the groundwater in both instances is either by diffuse intergranular flow or facilitated by faults resulting in submarine springs. The zone of the Upper Tertiary Carbonate Aquifer that receives the leakage is below the 0 mAHD contour and most likely is saturated with ocean water or a mixture of ocean water with groundwater in a lesser proportion.

5.3.2 Tyrendarra Sub-basin

The cross section of groundwater flow in the Tyrendarra Sub-basin also shows that interaquifer flow occurs readily in the form of local and intermediate flow systems; not just between the Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers but also with the Pliocene-Quaternary Volcanic Aquifer. Some of the thin surficial cover of the Pliocene-Quaternary Volcanic Aquifer could not be represented on the cross section. Flow to the southwest (denoted as \odot) is significantly present in this cross section and it relates to local southwestward flow towards the valley of Darlot Creek (Figure 4.5). Approximately 16 km along the section, the creek is crossed and this coincides with the first local discharge zone in the regional flow line. Other local discharge zones are geographically similar, although the one between the 60 and 50 mAHD contours coincides with the edge of a basalt flow. Local recharge zones include the topographic high at the beginning of the regional flow system and other local pressure highs, usually associated with volcanic cover. For example, the u-shaped 80 mAHD contour that crosses the Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers is the result of a groundwater mound in the vicinity of Mt Eccles. This particular recharge zone is important as it also serves as an intermediate and regional recharge zone.

The intermediate and regional flow systems are partly contained in the upper part of the Upper Tertiary Carbonate Aquifer, in which regional flow is seaward and the occurrence of the 0 mAHD isopotential contour at the ocean edge indicates coastal discharge and some diffuse submarine disharge. The Clifton Formation of the Upper Tertiary Carbonate Aquifer receives recharge through the overlying marl of the Upper Tertiary Aquitard but also benefits from the groundwater high marked by the u-shaped 80 mAHD contour. These findings agree well with those of Bennetts (2005) who asserted that significant interaquifer leakage involving the Pliocene-Quaternary Volcanic, Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers occurs in this region, ultimately recharging the Clifton Formation.

This cross section originates in the area at the edge of the Dundas Plateau where there are small exposures of the Lower Tertiary Sandy Aquifer. This is the main recharge zone for the Lower Tertiary Sandy Aquifer, and subsequently, the Upper Cretaceous Aquifer. However, some of the recharge does enter the aquifers via the thin overburden of other





aquifers or aquitards. Therefore, the quality of the deep groundwater is vulnerable to changes in the surficial system and changes in river water quality at this location. However, not shown in the cross section are the volcanic pipes of nearby eruption centres, e.g. the scoriaceous centre of Mt Napier or that of Mt Eccles and its neighbours further along the flow path. Although the vertical gradients shown in much of the cross section are upwards, due mainly to the location of the cross section near the incised valley of Darlot Creek, at the location of the eruption centres there is the potential for recharge to the deep aquifers. Approximately 30 km along the cross section the connection between the Lower Tertiary Sandy and Upper Cretaceous Aquifers is severed by the Lower Tertiary Aquitard. However, further interaquifer flow may occur between these formations closer to the coast where the aquitard thins again.

Upward hydraulic gradients from the Upper Cretaceous Aquifer are not observed in the onshore section but more data would be required to constrain this better. The Lower Tertiary Sandy Aquifer is subject to an upwards gradient at the coast. However, the horizontal hydraulic gradient and the thick confining layer of the Upper Tertiary Aquitard probably act to restrict vertical leakage until further along the flow path where upwards gradients presumably become stronger (c.f. Figure 4.16). Discharge from the Lower Tertiary Sandy Aquifer in this area appears to be further offshore than, for example, the scenario envisaged for the Gambier Sub-basin above. Both cases fit well with the conclusions of Chapter 4.

5.3.3 Port Campbell Sub-basin

The mainly southward and southwestward groundwater flow in the Port Campbell Subbasin (Figure 5.4) originates at the watertable divide between in the Port Campbell Subbasin and the Corangamite region, where the Pliocene-Quaternary Volcanic Aquifer is thick, at a high elevation and is directly overlying the Upper Tertiary Aquitard. The local flow systems are initially within these two hydrostratigraphic units and then encompass the Upper Tertiary Carbonate Aquifer also. Regional flow systems have developed in the upper and lower parts of the Upper Tertiary Carbonate Aquifer, the Lower Tertiary Sandy Aquifer and in the Upper Cretaceous Aquifer.

A zone of recharge exists a the northern edge of the cross section and persists along the flow line to the point of a depression (120 mAHD contour) which defines a local discharge zone associated with the cross cutting Mt Emu Creek. Another recharge zone, of local and regional importance, is represented by the concentric 130 and 140 mAHD contours approximately 5 km down gradient from the creek location. These isopotentials outline a mound in the watertable associated with Mt Noorat, a nested maar just north of Terang. As with the Tyrendarra Sub-basin, there is a local discharge zone close to the coast (below



Figure 5.4. Cross section A, showing distribution of hydraulic head and diagrammatic indication of groundwater flow along a regional flow line, within the hydrostratigraphic units of the Port Campbell Sub-basin.

the 80 mAHD isopotential) that coincides with the edge of a basalt flow of the Pliocene-Quaternary Volcanic Aquifer. Finally, the local flow system within the Upper Tertiary Carbonate Aquifer discharges at the coastline, implying that deeper, more regional flow discharges in the shallow submarine zone.

Intermediate and regional flow systems of this cross section generally originate at the Mt Noorat mound, although some intermediate flow within the Upper Tertiary Aquitard originates north of the Mt Emu Creek depression. Recharge to the regional flow systems of the Clifton Formation, the Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer occurs via diffuse leakage. However, it also probably occurs via faults and volcanic pipes not shown in the cross section. There are a large number of maars in the vicinity of Terang, e.g. Lake Keilambete or Lake Elingamite (Figure 4.5). These geological features could provide the pathways for leakage from the overlying units.

Discharge from the deep aquifers, as with the other sub-basins, is offshore via upwards leakage and cyclic flow with sea water that has infiltrated the aquifers.

5.4 Aquifer response to stresses

Up to this point, the current physical hydrogeology has been analysed and described as best as possible. Although steady-state conditions are assumed for the application of Darcy's Law (Chapter 4), the aquifers are known to be in a state of transient response to stresses and changes in boundary conditions. This means that the hydraulics must have been different in the past and will change in the future.

The importance of understanding past responses is that they provide the model for prediction of future responses to anticipated stresses such as increased demand on water resources or climate change. Another anticipated stress could include the further and ongoing development of oil and gas extraction in the basin, offshore and onshore. This particular issue has resulted in significant changes in the Gippsland Basin (refer to Hatton *et al.*, 2004), which is the same age as the Otway Basin. The more advanced stage of development of the hydrocarbon industry in the Gippsland Basin (Mehin & Link, 1994) may provide an insight into the future of the groundwater system behaviour of the Otway Basin, should development there continue and intensify. Towards a better understanding of the state of transience of the aquifers, the past changes in hydraulic head have been analysed.

The results of the hydrograph trend analyses are summarised in Figure 5.5, which shows the stretch of each bore's hydraulic head record over the years and the overall representative trend in hydraulic head (m/yr). Only eligible records are presented in this graph and records with a trend of -0.04-0.04 m/yr are not represented as they are



Figure 5.5. Rate of change of hydraulic head in bores versus dates of records. Bores from all aquifers are included. Bores that exhibited trends of -0.04–0.04 m/yr are not included as the trend is considered to be neutral.

considered to be neutral or stable trends. Initial observations are that falling trends are more numerous and reach greater magnitudes than rising trends. Also, that greater rates of change are associated with shorter records; in particular, short records in the late 1980s are mostly falling trends (Figure 5.5).

The distribution of falling and rising trends amongst the hydrostratigraphic units is summarised by the histogram in Figure 5.6. It should be noted that the neutral or stable trends, those between -0.04–0.04 m/yr, are depicted as their calculated values in this diagram. The majority of all records showed falling trends, as already gleaned from Figure 5.5. Further, each aquifer has its peak or major distribution in the falling (negative) range. The Upper Tertiary Carbonate Aquifer and the surficial aquifers, namely the Pliocene-Quaternary Volcanic Aquifer and the Plio-Pleistocene Heterogeneous Aquifer, have a peak distribution in the range of -0.05 - -0.1 m/yr. The Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer have a peak distribution in the -0.05-0 m/yr range. This could indicate that the shallow and deep aquifers respond differently to stresses and changes or that the factors influencing them are different. Another implication is that there could be the development of a steepening of falling trends and the shallower aquifers are further along in this development than the deeper aquifers. Conversely, the aquifer with the most records of rising hydraulic head is the Lower Tertiary Sandy Aquifer. As will be seen



Figure 5.6. A histogram showing the number of bores screened in different aquifers that record rates of change in hydraulic head over time. Italisied numbers are the number of all bores with a given rate of change in head.

below, the rising and falling trends within each aquifer are spatially distinct.

Although very short records were discarded from the dataset, the remaining short records had a tendency to include some high magnitude falling trends. This situation is depicted more clearly in Figure 5.7 where the length of the hydrograph record is plotted against the rate of change in head. Some of the trends with short record lengths must be treated with caution as they could be demonstrating a response on an interannual scale, rather than an interdecadal scale. However, the majority of the trends, whether rising, neutral or falling, are derived from records more than a decade in length (Figure 5.7).

Lastly, the hydrograph analysis results were plotted on maps for spatial analysis (Figures 5.8, 5.9 and 5.10). The maps show the hydrograph trends for bores screened in the surficial aquifers, the upper Tertiary units and the lower Tertiary/upper Cretaceous units; i.e. the shallow, mid-level and deep systems respectively. The definition of the shallow watertable system depicted in Figure 5.8 includes all bores screened in the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers and also encompasses eligible bores screened in any aquifer within 30 m of the ground surface. The categories of falling and rising trends used in the maps are the same categories used in Figure 5.7.



Figure 5.7. Length of data records of hydrographs plotted against the magnitude of the long term trend observed in the hydrograph. Negative trends are falling head levels and positive values are rising levels. Bores that exhibited trends of -0.04-0.04 m/yr are depicted in grey as the trend is considered to be neutral.

In some cases, unconfined, semi-confined and confined aquifers' water levels are presented in the same map. At these times the quantification of the rates of change in head are noted with respect to the nature of the aquifer. It is known from the difference between storage coefficients and specific yields that confined and unconfined aquifers respond differently to extraction. Accordingly, in response to stresses of equal magnitudes, rates of change in head may be greater in confined aquifers than in unconfined aquifers.

5.4.1 Shallow System

The map of the shallow aquifer bores (Figure 5.8) shows that a fall in the watertable has been occurring in much of the Otway Basin in the last three decades or more. The groundwater flow divides shown on this map were taken from the watertable map (Figure 4.4). The only area which shows stability of the watertable is the Gellibrand River region where, although one bore shows a falling trend and another a rising trend, the majority of the bores show no change in head. Those bores are mostly screened in the Lower Tertiary Sandy and Upper Cretaceous Aquifers.

North of the Upper Wannon region, near the head waters of the Glenelg River on the Dundas Plateau, watertable trends are mixed but mostly consist of strongly rising trends.





These bores are mostly screened in a laterised sand of the Plio-Pleistocene Heterogeneous Aquifer but some may be tapping the underlying fractured rock aquifers that are considered basement for the Otway Basin. The duration of these bores' records is short, either from 1991 to 1996 or from 1993 to 1997.

The remaining sub-basins, including the Upper Wannon region (Section 4.3), present a majority of falling trends with a few records of stable or rising watertable levels. The highest rates of watertable decline are recorded near the state border in the Gambier Subbasin and in the south of the Barwon Downs Sub-basin. Near the border, the bores are screened in the Plio-Pleistocene Heterogeneous Aquifer, the Upper Tertiary Carbonate Aquifer (including the Clifton Formation), and the Upper Tertiary Aquitard. In the south of the Barwon Downs Sub-basin, the bores tap the Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer and those in the north are screened in the Pliocene-Quaternary Volcanic and Plio-Pleistocene Heterogeneous Aquifers. The Corangamite region bores and the Upper Wannon region bores are mostly screened in the Pliocene-Quaternary Volcanic Aquifer. The bores in the Tyrendarra Sub-basin and the Port Campbell Sub-basin are screened in all the surficial aquifers.

5.4.2 Mid-level System

The changes in hydraulic head of the Upper Tertiary hydrogeological system, i.e. the Port Campbell Limestone, Gambier Limestone and Clifton Formation of the Upper Tertiary Carbonate Aquifer and the Gellibrand Marl of the Upper Tertiary Aquitard, have been mapped (Figure 5.9). The hydrograph trends for the Gambier Sub-basin are either falling or stable. Most of the Gambier Sub-basin bores are screened in the Gambier Limestone but some are in the deeper Upper Tertiary Aquitard and the underlying Clifton Formation.

In the northwest of the Tyrendarra Sub-basin, near the town of Condah and the groundwater flow divide with the Gambier Sub-basin, there are numerous strongly falling trends. These bores are in the Clifton Formation and the Gellibrand Marl; the bore with the mildest rate of decline taps the Port Campbell Limestone. To the east-southeast, the two stable trends on the Shaw River are from bores also screened in the Port Campbell Limestone.

In the coastal areas of the Tyrendarra and Port Campbell Sub-basins, there are many records of variable trends. All the bores in this region are screened in the Port Campbell Limestone and all but 3 have records that only span 5.5 years: from late 2000 to mid 2006. The three long records of this area each span more than 2 decades; one shows a fall of -0.07 m/yr and the other two show stable hydraulic head levels. As the aquifer is unconfined in this region, the hydrographs are subject to strong seasonal fluctuations. On one hand these variable results could indicate that localised controls on influx and outflux





are themselves variable and indeed important. Alternatively, the seasonal fluctuations may be unduly influencing the trends calculated from the short records. Ultimately, the results are inconclusive.

In the north of the Tyrendarra Sub-basin there are a few records from each of the hydrostratigraphic units, all showing declines in head levels (Figure 5.9). The Corangamite region and the Barwon Downs Sub-basin show trends that are either falling or stable and the bores represent the Gellibrand Marl or the Clifton Formation.

5.4.3 Deep System

The map in Figure 5.10 shows the trends for the Lower Tertiary Sandy and Upper Cretaceous Aquifers. This is labelled the deep system although the aquifers are exposed and shallow in the far east of the basin. Records are sparse in the Gambier and Tyrendarra Sub-basins but show either falling or stable trends.

Contrastingly, the majority of the Port Campbell Sub-basin contains bores with rising head levels. However, towards the east of this sub-basin, near the upper reaches of the Gellibrand River, a dominance of stable head records is apparent. In the very northeastern corner of this Sub-basin the bores' hydrographs show falling trends. Rather than these zones of differing changes in head being separated by groundwater flow divides, they are defined by structural boundaries. The stable zone is separated from the rising zone by the basement outcrop at the crest of the Barongarook High. The transition from the stable zone to the falling zone is the groundwater flow constriction at the Gellibrand saddle (Section 4.8), created by the truncation of the aquifers and the narrowing of their lateral extent.

The hydrograph trends of the Barwon Downs Sub-basin are almost exclusively negative (Figure 5.10). However, some stable or rising trends do exist, especially along the northern part of the divide with the Port Campbell Sub-basin. These trends are structurally separated from the falling trends by the Barongarook High.

5.4.4 Stresses

A decline in water levels in the Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers of the Gambier Sub-basin have been identified in the past (e.g. Ingram *et al.*, 1989) and are currently recognised as a management issue (Border Groundwaters Agreement Review Committee, 2006, 2008). The 2006 report identified as the causes of the decline in the Upper Tertiary Carbonate Aquifer head levels: groundwater extraction; reduced recharge linked to decline in precipitation; and increased transpiration from plantation expansion. The 2008 report noted that further investigation is needed to ascertain





the cause of the falling head levels in the Lower Tertiary Sandy Aquifer, but that it would appear that the "current mix of land use and groundwater extractions is clearly out of balance and is not sustainable in the longer term". All these factors, along with the influence of changes in sea level and the development of the basin's hydrocarbon industry, are explored below.

5.4.4.1 Groundwater extraction

Extraction from the Plio-Pleistocene Heterogeneous and the Pliocene-Quaternary Volcanic Aquifers is common throughout the basin. It is generally for stock and domestic purposes with some extraction for irrigation and dairy wash. Significant extraction from the Pliocene-Quaternary Volcanic Aquifer occurs in the Warrion district (north of Colac; Figure 5.8). In addition, some small towns obtain their municipal supply from that aquifer, namely: Caramut, Mortlake and Skipton (Leonard, 2003; Figure 5.8). From analysis of the hydrographs in the surficial aquifers, is can be seen that a large number of bores present declining trends (Figure 5.8). The declining trends are partially due to the groundwater extraction, although some influence from other factors will also be revealed as important.

The bores of the Plio-Pleistocene Heterogeneous Aquifer in the Dundas Plateau region (near Coleraine, Figure 5.8) present some strongly rising and some strongly falling trends. Groundwater extraction is not intense in this area and therefore not likely to be the main causal factor of the falling trends.

Groundwater from the upper part of the Upper Tertiary Carbonate Aquifer (the Port Campbell Limestone and the Gambier Limestone) is used to supply many domestic and agricultural users throughout the entire Basin and is also the municipal supply for the township of Warrnambool, the fastest expanding regional centre of the Victorian section.

Extraction from the Upper Tertiary Carbonate Aquifer in the Border Zone of the Gambier Sub-basin is especially intensive in the areas closest to the coast (Border Groundwaters Agreement Review Committee, 2008). That report classified parts of the Upper Tertiary Carbonate Aquifer as over-allocated, indicating that the declining head levels in the coastal part of the aquifer (Figure 5.9) are at least partly a result of abstraction. However, when the rates of decline in the map are inspected, it is apparent that there are faster rates of decline in the north of the Gambier Sub-basin than in the south (Figure 5.9). This pattern contradicts that of groundwater usage (which is more intense in the south) and therefore indicates that another factor is jointly responsible for the change in hydraulic head over time.

The Upper Tertiary Carbonate Aquifer is also subject to significant groundwater extraction in the coastal regions around Port Fairy and Warrnambool. However, rates of change there are mixed (Figure 5.9). This is an area where the aquifer is close to the surface and so is interacting closely with recharge water, river water (e.g. the Hopkins River, the Moyne River, and others) and seawater. Hence, the local hydrological balance for the aquifer is subject to changes in any of those bodies of water. When looking at rates of change in the other unconfined bores in the region, Figure 5.8, rates are similarly mixed. This confirms that local surface processes are indeed influencing the Upper Tertiary Carbonate Aquifer. Groundwater extraction is one such local influence at locations with declining rates of change in head.

The groundwater of the Clifton Formation is utilised extensively for irrigation in the Condah region (Bennetts, 2005). The cluster of bores screened in the Clifton Formation and the Upper Tertiary Aquitard in the Condah region (Figure 5.9) illustrate the effect of the extraction with strongly declining hydraulic head levels. This is the first case discussed that involves a completely confined aquifer, which one might reasonably assume would have a storativity value similar to the Lower Tertiary Sandy Aquifer (Table 4.1). This value being significantly lower than the specific yield values of the unconfined aquifers of the basin (Table 4.1). Accordingly, the rates of falling pressure are more pronounced in the Clifton Formation than the rates of decline in the surficial, unconfined aquifers (c.f. Figure 5.8).

There are other regions where there are records of declining hydraulic head in the Clifton Formation, and also in the Upper Tertiary Aquitard. However, these layers are not subject to groundwater use in these particular regions, which are the north of the Port Campbell Sub-basin, the Corangamite region and the south of the Barwon Downs Sub-basin (Figure 5.9; c.f. Figure 4.13). As no groundwater is currently pumped from these aquifers at these locations, the trends are likely to be sympathetic reactions to declines in overlying or underlying units. The first two regions are likely to be experiencing reduced recharge due to changes in the water balance in the overlying Pliocene-Quaternary Volcanic Aquifer (Figure 5.8), which in turn are partly the result of abstraction. Contrastingly, the decline in the southern part of the Barwon Downs Sub-basin is most likely linked to extraction from deeper units: the Lower Tertiary Sandy and Upper Cretaceous Aquifers (see below).

The groundwater from the Lower Tertiary Sandy Aquifer is used by Wannon Water to supply numerous towns in western Victoria, including Port Campbell, Timboon, Peterborough, Heywood, Dartmoor, Portland and Port Fairy (Figure 5.10). Town supplies in South Australia are also supplied by the Lower Tertiary Sandy Aquifer (e.g. Naracoorte; O'Driscoll, 1960). In addition, the Lower Tertiary Sandy and Upper Cretaceous Aquifers are both tapped in the Barwon Downs Sub-basin (north of Gellibrand) to augment the water supply for Geelong. The Lower Tertiary Sandy Aquifer is widely used for water resources due to the low salinity of the groundwater (TDS range of 300-5600 mg/L) and also due to the aquifer accessibility in the eastern parts of the basin (Leonard, 2003). Bore fields from this aquifer can discharge up to 18 ML/day and abstraction is known to have created cones of depression (Skidmore, 1992; SKM, 2005).

Prior to 1957, the Portland water supply was taken from the Upper Tertiary Carbonate Aquifer (bore screen depth was 66–168 m; Shugg, 1981). In 1972, *Portland 11*, a bore more than 1,200 m deep, was being used to extract the town supply at a rate of 6.15 ML/day from the Lower Tertiary Sandy Aquifer (Shugg, 1976). Currently, the Portland water supply is still pumped from the Lower Tertiary Sandy Aquifer and consists of approximately 3.5 ML/day in winter and 12 ML/day in summer (Wannon Water).

Records of hydraulic head for the Lower Tertiary Sandy Aquifer in the Portland area show that levels have been falling, on average, by less than 0.2 m/yr (Figure 5.10). However, as mentioned previously, the digital records used in this study are not all-inclusive. When comparing singular measurements of head from the past, faster rates of decline become apparent. In the late 1950s to early 1970s, the potentiometric surface near Portland was over 20 mAHD, and some levels were even over 30 mAHD (Shugg 1976). Today it is around 0 mAHD (Figure 4.14). From these values an approximate range of rates can be calculated and equates to -0.75--0.45 m/yr. These falling head rates are significantly more negative than those depicted in the map and illustrate the cone of depression that has developed in response to the extraction at Portland. Similarly, the potentiometric surface of the Lower Tertiary Sandy Aquifer near the coastal town of Port Fairy also shows declining trends due to the extraction there.

However, the same cannot be said for the negative rates of change in the coastal area near Nullawarre (at the divide between the Tyrendarra and Port Campbell Sub-basins; Figure 4.5). There is no significant extraction from the Lower Tertiary Sandy Aquifer in the Nullawarre region. For the same reason, declining rates in the Gambier Sub-basin and the recharge zone of the Tyrendarra Sub-basin cannot be attributed to extraction.

The Port Campbell Sub-basin, including the Gellibrand River region, contains mostly positive or stable trends for the Lower Tertiary Sandy Aquifer, despite extraction at Port Campbell and Carlisle River (12 km southwest of Gellibrand). The annual extraction at Carlisle River is approximately 1,800 ML/yr (Wannon Water) and is piped to nearby towns to augment supplies. However, in the eastern corner of the Port Campbell Sub-basin, just south of the Gellibrand Saddle, the negative change in hydraulic head in the Lower Tertiary Sandy and the Upper Cretaceous Aquifers is linked to the extraction in the Barwon Downs Sub-basin to the north.

The Barwon Downs bore field, in the parish of Gerangamete just east of Kawarren, produces groundwater from the Lower Tertiary Sandy and Upper Cretaceous Aquifers to
supplement the Geelong water supply (refer to Petrides & Cartwright, 2006, for details). The water is extracted from the Dilwyn Formation, sandy facies of the Mepunga Formation and from the Pebble Point Formation. In this region the Lower Tertiary Aquitard, which usually separates the aquifers, is absent, enabling local hydraulic interconnection. As a result of this extraction, strongly negative trends in head have been recorded over time in the Aquifers themselves (Figure 5.10) and also in the overlying Gellibrand Marl and Clifton Formation (Figure 5.9). Despite the long term effect the pumping has on the shallower units, there is no strong correlation between the timing of pumping from the deeper aquifers and any record of bore drawdown in the overlying aquifers and aquitards (Petrides & Cartwright 2006).

The long term decline as a result of extraction in the deep aquifers at Barwon Downs was anticipated in the hydrogeological assessments around the time of the development of the bore field (Leonard *et al.*, 1983). The effect of the pumping was also anticipated to reverse hydraulic gradients, such that any southwestward flow from the Barongarook High over the Gellibrand saddle to the Port Campbell Sub-basin would be rotated to flow north to the Barwon Downs Sub-basin (Leonard *et al.*, 1983). This has clearly not occurred, as the groundwater divide between the sub-basins still resides significantly north of the Gellibrand Saddle (Figure 4.14). What is interesting to note, though, is that the stress of the pumping has migrated over the groundwater divide and negative trends are observed in the eastern corner of the Port Campbell Sub-basin (Figure 5.10).

5.4.4.2 Hydrocarbon development

As mentioned previously, the extraction of hydrocarbons in the Gippsland Basin, mainly offshore, has contributed to major changes in the hydraulic balance of the Latrobe Group aquifer system (Nahm, 2002; Hatton *et al.*, 2004). The extraction of the hydrocarbons and incident extraction of formation water has led to a significant depressurisation of the aquifer that has been transferred to onshore regions.

The Otway Basin is generally less developed than the Gippsland Basin; it has had several commercial hydrocarbon discoveries, but most are gas, with few strong oil shows (Mehin & Link, 1994). When compared to the combined oil and gas resources in the Gippsland Basin, the discovered reservoirs of the Otway Basin are much smaller. Part of the reason the Otway Basin is less productive than the Gippsland Basin is that the rocks of the Otway Basin that are contemporary to the major source rocks in the Gippsland Basin, i.e. the Sherbrook, Wangerrip and Nirranda Groups (Figure 2.4), are mostly thermally immature (O'Brien & Thomas, 2007). Hence, exploration in the Otway Basin, which started onshore in the 1860s (Sprigg, 1986) and offshore in 1959 (O'Brien & Thomas, 2007), has not always been as active as that in the Gippsland Basin. However, aided by high quality seismic interpretations in the 1990s (Sarma, 1996) and later 3D seismic interpretation, exploration and production in the Otway Basin have seen a resurgent interest in recent years. New exploration permits are still being awarded at present and prospective hydrocarbon resources have been estimated to include up to 3.6 trillion cubic feet of yet undiscovered gas (O'Brien & Thomas, 2007). Therefore, production may be expected to increase into the future and investigations into the future impacts on the groundwater system are warranted.

O'Brien and Thomas (2007) provide an appraisal of the current state of exploration and production in the basin as well as a summary of the different source possibilities for the hydrocarbons. The majority of commercial finds have been gas fields whose source rocks are generally the Eumeralla Formation. Some small oil reservoirs have been found, mostly in South Australia, and their source is usually strata older than the Eumeralla Formation (e.g. the Laira Formation, the Pretty Hill Sandstone or the Casterton Formation; Figure 2.4; O'Brien & Thomas, 2007). Therefore, traps are situated in the strata overlying the Eumeralla Formation. The Waarre and Flaxman Formations, despite being themselves possible source rocks, contain some of the gas fields. Hydrocarbons are sometimes mobilised through mixing with magmatic CO_2 , or can be entirely flushed from traps by the latter. Migration occurs via faults (Mehin & Link, 1994).

Hydrocarbon development at onshore gas fields (e.g. Iona, which is now also an underground gas storage site) and offshore (e.g. the Henry and Casino gas fields) in the region of Nullawarre is most likely contributing to the decline in pressure in the Lower Tertiary Sandy Aquifer in that region. Further investigation would be required to confirm this.

In addition to the gas injection for storage at Iona, the Nullawarre–Port Campbell region is the location of a pilot project to trial geosequestration of CO_2 . The project is a AU\$40 million endeavour of the Australian and Victorian governments, along with other partners. The CO_2CRC (co-operative research centre) has overseen the project and has recently announced the successful sequestration of more than 50,000 tonnes of CO_2 to a depleted gas field in the Waarre Formation. The gas was extracted from a separate field via Butress-1 (Xu *et al*, 2007) and injected via a new well (CRC-1). The migration of the gas is monitored 300 m from CRC-1 at Naylor-1 and results are compared to modelled scenarios (Ennis-King & Paterson, 2007). In addition, potential leakage is monitored by groundwater, soil and atmospheric monitoring (A. Hennig pers. comm. 2007). No leakage has been detected at this stage and monitoring will continue until at least 2010. Ongoing monitoring of groundwater may result in an increased understanding of this project's impact on the groundwater resources of the basin.

5.4.4.3 Climate change & Sea level change

In Chapter 3 the current climate system is outlined, along with the past changes in climate and sea level with special attention to the effect of glaciation. It was found from palaeoclimate evidence that the climate of the Otway Basin region was wetter than today from 2 ka to about 200 years ago. In the last 200 years the climate has become drier (i.e. has experienced a lowering of the P/E ratio). This trend is confirmed for the last 100 years by historic meteorological observations of temperature and rainfall in the region and has been exacerbated by anthropogenic increases in greenhouse gas concentrations in the atmosphere and other factors.

Sea level during the last glacial maximum was approximately 120 m lower in the region of the Otway Basin. It rose with deglaciation and reached the current level at 6 ka, resulting in an average transgression rate of 0.0086 m/yr. After 6 ka, the sea level is inferred to have been relatively stable until about the last 150 years. In the last century (1900–2000), sea level has risen, on average, at a rate of 0.0017 m/yr (IPCC, 2007).

The length of the lag involved in the groundwater system's response to sea level changes is important because of the time frames discussed above. If the lag is greater than 6,000 years or less than 150 years, then it is reasonable to expect an aquifer response to rising sea levels, i.e. a rise in hydraulic head and/or a shallowing of coastward hydraulic gradients. If the lag is between these estimates, then it may be reasonable to assume that sea level changes are not currently affecting the hydraulic head regimes of the aquifers. A lag also applies to the system's response to climate change. The recent drying of the climate ultimately leads to a decrease in recharge and an increase in discharge via evaporation, which together cause a decrease in hydraulic head. Hence, depending on the lag time, the two factors of climate change and sea level change may have opposing effects on the aquifers.

The difficulty lies in separating out the interfering signals of the various stresses. By discounting for a moment the recent anthropogenic changes to the groundwater system via extraction (see above) and land use change (see below), one may consider which of the two factors, sea level change and climate change, may be more important. If climate alone is the major factor controlling the hydrologic balance, then the changes in the surface water regime must lead those in groundwater levels. Put differently, surface water drainage (e.g. lake levels) respond quickly and sensitively to changes in the P/E ratio, whereas the watertable response to such changes is subject to a lag. Alternatively, if the P/E ratio were to remain relatively constant while the sea level changed significantly, then the hydraulic connection between the ocean and the aquifers would induce pressure changes in the groundwater system which would in turn control lake levels.

the rate and timing of changes in lake levels and groundwater levels, it may be possible to determine the main driving factor behind the changes. It is also reasonable to extend this idea to monitoring the different responses in shallow aquifers (more quickly responsive to climate change) and deep aquifers.

The only surface water records suitable for comparing with groundwater records are lengthy ones, as short records are too susceptible to annual- or decadal-scale fluctuations in climate. Long records of observations of lake levels exist for several lakes in the Otway Basin. Some observed rates of change (with the record spans provided in parentheses) are: -0.14 m/yr for Lake Keilambete (1859–1990); -0.08 m/yr for Lake Gnotuk (1964–1988); and -0.17 m/yr for Lake Bullen Merri (1859–1990; Jones *et al.*, 2001). In comparison, a rate of -0.08 has been estimated for Lake Keilambete for the longer period of 1750–1950 A.D. based on reconstructed lake levels from palaeoclimate proxies (Bowler, 1981). Simply by the fact that rates of lake level change are negative for the last 100 to 150 years, the influence of recent sea level rise on shallow groundwater trends is ruled out. Instead, the negative trends of the lakes are linked to climate change. The lake records span a time frame before the intensification of groundwater use in the region and therefore indicate that the pre-existing factor of climate change is indeed valid. Extraction of groundwater remains, however, a conflicting and potentially mutually amplifying factor.

The average rates of lake level decline agree well with the rates of decline in the Pliocene-Quaternary Volcanic Aquifer of the Corangamite region (Figure 5.8). However, some stable trends do exist in the aquifer records, indicating that, groundwater changes may be lagging behind surface water changes. In addition, the shallow aquifers (the Pliocene-Quaternary Volcanic, Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers) have peak distribution of bores at a greater falling rate than the deeper aquifers, possibly indicating a "top-down" mechanism of increasing rates of decline (Figure 5.6). Again, these observations support the conclusion that climate change is important in driving declines in all bodies of water.

However, the reduction in net recharge over time in the surficial aquifers is not due to a decrease in actual annual recharge (Section 4.4). Rather, an increase in discharge is responsible and this could be either increased extraction or greater evapotranspiration. A contribution of both factors is expected to be important in the hydrological balance of the Otway Basin, specifically, in the case of the volcanic lakes (Adler and Lawrence, 2008). This conclusion is also valid for the unconfined aquifers in other regions of the basin, especially in the recharge zones of the Tyrendarra and Gambier Sub-basins (Figures 5.8 and 5.9).

The abundance of stable watertable levels in the Gellibrand River region indicate a system close to steady state conditions (Figure 5.8). The groundwater in this area is not

greatly exploited and is structurally isolated from extraction from the Pliocene-Quaternary Volcanic Aquifer in the north and from the Lower Tertiary Sandy and Upper Cretaceous Aquifers in the northeast. In addition, it is not confined and is fairly distal from the ocean and so sea level change can be largely ruled out is an influence. The result is that climate and land use changes are the only remaining factors that could affect the water balance of this region of the watertable. The region has experienced some change in land use with European settlement, however, the hilly topography meant that many inaccessible regions where not deforested. Therefore, the question of why climate change has had no notable effect on the hydrologic balance in this area remains. An explanation may lie in the fact that the precipitation in excess of evaporation in this area is higher than the rest of the basin, due to the orographic rainfall high over the Otway Ranges. The high humidity of the region may be slowing the effect of increased evaporation and allowing the groundwater system to remain in equilibrium.

In the remaining parts of the Port Campbell Sub-basin, where the deeper aquifers become confined, they continue to show stable trends in hydraulic head. However, some bores exhibit significant increases in hydraulic head. The areas where these bores are situated are sometimes close to the coast and are certainly away from the unconfined recharge zone of the Gellibrand River region. The increase in head in these bores may be an hydraulic response in the confined aquifers to sea level rise since the last glacial maximum. In other words, the onshore and offshore hydraulic head in the aquifer was lower in the past, due to the low sea level, and is still adjusting to the changes of the Holocene (refer to Section 4.8.4). Love *et al.* (1994) agreed that the past regime was different and that this caused discrepancies between groundwater ages measured by Darcy's Law and those by isotopic dating. However, they indicated that the Lower Tertiary Sandy Aquifer is in equilibrium with the current sea level.

The falling levels in the recharge zones of the Lower Tertiary Sandy Aquifer in the Gambier and Tyrendarra Sub-basins are attributed entirely to reduced. Both climate change and extraction from overlying aquifers are factors in this reduced recharge. Therefore, although extraction from the aquifer in those regions does not occur, extraction from other aquifers still plays a part in the hydrologic balance of the deeper unit.

5.4.4.4 Land use change

It has been shown by several authors that land use, specifically vegetation cover and related evapotranspiration, is an important factor in recharge of the Otway Basin (e.g. Holmes & Colville, 1970a & 1970b; Leaney & Herczeg, 1995). In turn, changes in recharge over time will have an effect on long term hydraulic head levels. There are many scenarios to be considered in the Otway Basin in terms of land use change, however, only two gen-

eral scenarios involving vegetation will be addressed below. They are: 1) the decrease in transpiration due to removal of native vegetation during European settlement; and 2) the increase in transpiration from expansion of *Pinus radiata* and *Eucalyptus globulus* plantations. A third scenario of land use change important in the Otway Basin is anthropogenic changes to the natural drainage systems, reducing the number of wetlands.

The widespread practice of land clearing by European settlers in south eastern Australia is well documented in the literature and this practice was pervasive in much of the non-mountainous areas of the Otway Basin (Love *et al.*, 1993). However, it should be noted that the original native vegetation in parts of the basin, specifically the basalt plains, was not dense woodland or bush, but open grassland (Dahlhaus *et al.*, 2008) with a similar transpiration rate to pasture. In the Murray Basin, the documented result of land clearing is that the potential for recharge has generally been increased, as native vegetation was replaced with pasture and the transpiration rate dropped accordingly. The watertable levels have risen steadily in some areas, becoming problematic by mobilising and concentrating salt in the groundwater (Herczeg *et al.*, 2001). This well established process has not been observed in the Otway Basin proper, and few locations have an abundant distribution of rising head levels (Figures 5.8, 5.9 and 5.10). In fact, the areas of the Otway Basin where groundwater and surface water salinisation can be problematic, e.g. the Corangamite region, show declining water levels (this study; Dahlhaus *et al.*, 2008).

However, one area that does have rising head levels is the Dundas Plateau, where there are several bores screened in the Plio-Pleistocene Heterogeneous Aquifer and the fractured bedrock (Figure 5.8). These trends may have been influenced by local changes in land use, however, their records are restricted to the period from 1991 to 1997 and increases may also relate to precipitation trends at the time (see Figure 3.6). In addition, the proximity of falling trends makes the results difficult to decipher and probably points towards localised factors. The Dundas Plateau has a similar geological framework to that of an area studied by Bennetts *et al.* (2006) to the east, near Glenthompson (refer to Figure 4.5). Both regions have basement rocks near or at the surface acting as fractured-rock aquifers and a thin cover in some parts by volcanics and/or Quaternary sediment. It was found that the salinity of the Glenthompson area is primary and had been accumulating for most of the Holocene, but that in recent decades the rise in watertable due to land clearing has exacerbated the salination (Bennetts *et al.*, 2006). This indicates that similar processes may de influential in the Dundas Plateau. Salinity will be discussed in more detail in Chapter 6.

In recent years there has been a significant increase of land area occupied by private forestry in the Otway Basin, both in Victoria and in South Australia. There is a known correlation between forest cover and low effective deep drainage (Holmes & Colville, 1970a). A recent shift in land use means that changes to recharge may be anticipated immediately and into the future. The shallow unconfined aquifers are the most susceptible to the effect of plantations. However, the flow-on effects are regional because, as explained above, the deep aquifers often rely on diffuse leakage for their recharge. Areas of the Gambier Subbasin and the Tyrendarra are the most heavily forested and the effect is reflected in the prevalence of falling hydraulic head in these regions. The effect of increased transpiration from plantations combines with increased evaporation from climate change and increased demand on groundwater resources to create an unavoidable stress on the aquifers.

Perhaps exacerbating this problem are the changes to the natural drainage system in the last 100 years or so. As land use changed in the Otway Basin from sustainable practices, such as the eel farming community of Condah Swamp (Builth, 2009), to European agricultural practices, the landscape was changed, too. The swamps of the interdune swales of the Gambier Sub-basin presented a problem for European style farming practices. The swamps were systematically drained by artificial methods, perhaps the most extreme of which was the digging of a trench through the Woakwine Range called McCourts Cutting (refer to Murray-Wallace *et al.*, 1999, for location). The swamps, especially the ephemeral ones, are known have contributed seasonal recharge to the underlying aquifers and their drainage has no doubt contributed to the fall in hydraulic head in the aquifers, both unconfined and confined, of the Gambier Sub-basin.

More recently, alteration of the drainage network in the Corangamite region was enacted to protect farming land close to Lake Corangamite from flood waters. The diversion of Woady Yaloak River (Figure 4.5) from termination at the lake to instead discharge at the ocean occurred in the 1960s (Adler & Lawrence, 2008). This loss of inflow has contributed to the shrinking of the lake and concentration of its salinity, threatening the biodiversity that has won the lake Ramsar status (Timms, 2005). In light of the discussion above, other factors, including interaction with the groundwater system, are also important factors in the water balance for Lake Corangamite. Timms (2005) suggests that, to restore the former conditions of the Lake, better flood management is needed. However, given the interaction with the groundwater system proven here, it would seem that unmitigated watertable decline may undermine such efforts.

5.5 Synthesis

In this chapter, the examination of three hydraulic cross sections, created by transferring data from the potentiometric maps of Chapter 4, has confirmed several key mechanisms relating to vertical groundwater flow of differing scales.

First, all the sections confirmed interconnected nature of local- and intermediate-scale

flow systems within the surficial aquifers and the underlying units, be they the Upper Tertiary Carbonate Aquifer or the Upper Tertiary Aquitard. In the case of the Corangamite region, the aquitard underlies the Pliocene-Quaternary Volcanic Aquifer and creates a region where vertical drainage is inhibited. Local recharge zones are diffuse near the basin margins and point source at volcanic eruption centres. Local discharge zones are often creeks and rivers, even in the up-gradient zone of the flow path. This means streams receive baseflow at several locations. Local discharge can also occur at edge of a basalt flow of the Pliocene-Quaternary Volcanic Aquifer.

Second, the intermediate- and regional-scale flow systems are linked with these local systems and often share a recharge and/or discharge zone. Regional recharge can be linked to groundwater mounds from volcanic eruption centres or to high pressure head in the margin areas. Although faults and volcanic pipes are undoubtedly important pathways, the Clifton Formation and the Lower Tertiary Sandy Aquifer are also reliant upon diffuse leakage through the Upper Tertiary Aquitard for recharge. Also relevant to leakage is the Lower Tertiary Aquitard. The flow sections show that where the aquitard is thin or absent, interaction between the Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer is more viable. Lastly, the cross sections confirm the coastal and submarine discharge that was proposed for the Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers in Chapter 4 and highlight the differences between the sub-basins in this respect.

After analysing the hydrographs for 598 bores tapping the aquifers of the Victorian part of the Otway Basin, increasing, stable and decreasing trends over the last several decades have been interpreted as responses to specific stresses and/or changes in boundary conditions. A particularly important finding is that stresses can be transferred across groundwater flow divides. Physical or structural boundaries, such as the shallowing of depth to basement or the pinching out and truncation of aquifers, are better constraining bounds for the influence of stresses than groundwater divides.

Each of the aquifers is under stress from groundwater extraction in at least one region of the Otway Basin. Strongly declining trends in hydrographs have been identified in areas where there are no other known changes to boundary conditions (e.g. northeast corner of the Port Campbell Sub-basin) and are attributed to extraction. Extraction not only causes decline in the aquifer from which the groundwater is pumped, but also affects underlying and overlying aquifers and aquitards, as exemplified in the Condah region with extraction from the Clifton Formation and sympathetic decline in head of shallower units. This observation demonstrates induced leakage into the Clifton Formation and fits well with the interaquifer flow systems described above.

However, the effect of a drier climate for about the last 200 years are also evident in declining hydrograph trends and often a declining trend is due to a combination of climate

change and extraction. In the Corangamite region, the rate of decline of lake levels exceeds or is about the same as the rate of decline of the watertable. In addition, more bores in the surficial aquifers have rates of decline ranging from -0.1 to -0.05 m/yr than any other range, whereas the most popular range for the Lower Tertiary Sandy Aquifer is the more gentle decline of -0.05–0 m/yr. These observations indicate a "top-down" causal factor that is probably climate change. One area not affected up to now is the Gellibrand River region. The steady trends in shallow bores indicate that the relatively more humid microclimate of the Otway Ranges has protected the groundwater from a recent increase in evaporation.

Sea level is influential on the hydraulic balance of the Otway Basin, however, there is not a lot of evidence in existing hydrographs of the effect of the sea level rise in the last century. There are some rising trends in the Lower Tertiary Sandy Aquifer that may be the delayed response to transgression since the last glacial maximum.

The effect of hydrocarbon extraction is possibly seen in the Nullawarre–Port Campbell region in declining water level trends in deep aquifers. As production and exploration can be expected to increase in coming decades, more research into the repercussions for groundwater resources is necessary. Other expanding ventures in the Otway Basin, such as deep gas injection for storage, CO_2 geosequestration and geothermal exploration, should also be studied, as their affects on the system are currently largely unknown.

Land use change in the form of native vegetation clearing is deemed to have not caused much stress on the aquifers. The drainage of lakes and lagoons for the purpose of agriculture expansion or security has had a negative effect on the shallow groundwater system and poses a significant risk to groundwater dependant ecosystems. Expansion of plantations, notably in the Gambier Sub-basin, is likely to be contributing to the strongly falling watertable.

Chapter 6

Major & Minor Ion Hydrogeochemistry

6.1 Previous Work

Hydrochemically, the groundwater of the basin is highly variable, representing a suite of weathering reaction products. Johns (1968) divided the groundwater of the Lower Tertiary Sandy Aquifer into 4 classes using Na/Cl and Mg/Cl ratios. The distribution of the groundwater types indicated that some along-flow evolution occurred in the aquifer, however, sparse data prevented complete constraint of reactions. Processes that affected the groundwater chemistry, according to Johns included: cyclic input of Cl, cation exchange, water-rock interaction, and diagenetic bacterial reduction of sulphate (Johns, 1968; Johns, 1971).

A comprehensive study was conducted by Blake (1980) that concentrated on the structural geology and hydrogeology of the Lower Tertiary units in the Victorian section of the Otway Basin. Groundwater high in sodium bicarbonate was considered to be produced via feedback between silicate water-rock interaction and continual cation exchange and flow lines exhibiting chemical evolution were identified. Dolomite dissolution was also identified in the lower Tertiary aquifers (Blake, 1980). Areas of high chlorinity were identified over the Warrnambool high and were attributed to ineffective flushing of the region by meteoric inputs.

Since that time, several studies have concentrated on the chemistry and isotopic composition of the groundwater of sub-basins within the Otway Basin, such as the Gambier Sub-basin (Love *et al.*, 1992a; Love *et al.*, 1993; Leaney & Herczeg, 1995; Herczeg *et al*, 1997); the Tyrendarra Sub-basin (Bennetts, 2005); Port Campbell Sub-basin, (Duran, 1986) and the Barwon Downs Sub-basin (Leonard *et al.* 1983; Leonard & Lakey, 1989; Petrides & Cartwright, 2005). Maddocks (1967a, 1967b) studied the surface water chemistry in the Corangamite region. These and other works are drawn upon in the discussion to compare and contrast with results and conclusions from this study.

6.2 Methods

6.2.1 Data analysis methods

Two sets of publicly available data relating to the hydrogeochemistry were analysed. The first comprises the electrical conductivity (EC) records from all groundwater wells in the South Australian (Appendix I) and Victorian (Appendix H) sections of the Otway Basin. All these data are recorded as $\mu S/cm$, standardised to 20°C. Although multiple records exist for some boreholes, time series are not presented in this study. Records were divided according to the aquifer in which the borehole is screened (Appendices C & D). The most recent EC data for each location within each aquifer were used to create a map for that layer. The maps were created by inverse distance interpolation in GoCAD; the file is provided in Appendix J. The lithology of the surficial aquifers, the Plio-Pleistocene Heterogeneous Aquifer and the Pliocene-Quaternary Volcanic Aquifer, is very heterogeneous and the aquifers are not contiguous. Therefore they are too problematic to be represented separately in regional maps. For this reason, and to illuminate the interaction of these aquifers with the deeper aquifers where they outcrop, a watertable EC map was created from data from boreholes screened within 20 m of the land surface.

The second publicly available dataset that was analysed contains the concentrations of major anions and cations in groundwater from Victorian observation wells. The boreholes were classified according to the aquifers they tap using the database in Appendix C and subsequently records for the Pliocene-Quaternary Volcanic Aquifer, the Upper Tertiary Carbonate Aquifer and the Lower Tertiary Sandy Aquifer were retained. Mass to volume concentrations were converted to milliequivalents per litre (meq/L) and the relative proportions of cations and anions in solution were calculated (%). Results were analysed in comparison with overall salinity (as represented by $\Sigma_{cat} + \Sigma_{an}$ in meq/L) and relative position in regional flow paths. The latter is achieved by representation of salinity and chemical composition in cross sections. The data are presented graphically in the discussion and are provided in full in Appendix G.

Of the 6775 records of groundwater chemistry from Otway Basin aquifers in Victoria, 91% had a charge error balance $< \pm 5\%$. Hence, despite not having Br⁻ values, which are useful due to bromide's conservative nature, these records are a very valuable resource for understanding the evolution of the groundwater in the system.

These two data resources, the groundwater chemistry and EC, were combined to create

generalised cross sections of groundwater quality and chemistry in the different sub-basins. These cross sections are useful in describing the typical evolution of the groundwater along regional flow paths in the Upper Tertiary Carbonate Aquifer (Section 6.3.3) and the Lower Tertiary Sandy Aquifer (Section 6.3.5), although a lot of existing local variation can unfortunately not be represented on the regional scale. The cross sections are the same as those presented in Chapter 4, where the flow regimes for each are described (locations in Figure 2.3).

The electrical conductivity contours in the cross sections are derived from the EC distribution in the different hydrostratigraphic units as shown in Figures 6.6, 6.8, 6.14 and 6.16 and were contoured by hand. In some cases a detailed understanding of the lateral distribution of EC is not known and this creates uncertainty when transferring this data to cross section. The result is that the contours presented are preliminary attempts only and should be considered as interpretive.

The chemical composition charts in the cross sections were hand picked from the dataset of the Victorian water chemistry records and the chemistry results from this study to be representative of the aquifer in a particular part of the flow system. The relative abundance values represented in the charts were calculated from concentrations in meq/L, as described below. The proportion labelled HCO_3^- includes bicarbonate and carbonate concentration. It should be noted that significant variation from the typical compositions is observed at most locations.

6.2.2 Field and laboratory methods

This Chapter presents the major and minor ion hydrogeochemistry data whereas Chapter 7 deals with the isotope geochemistry. However, for the sake of simplicity, the field and laboratory methods for both groups of analytes are presented together below.

The field work for this project was conducted in several rounds (Table 6.1). In all, 177 boreholes and 24 surface water sites were visited throughout the Victorian and South Australian sections of the basin, although not all sites were sampled. The samples collected for various laboratory analyses are outlined in Table 6.1. In addition to sampling, various field methods, also outlined below, were used to gather hydrogeological information. The results from all field measurements and laboratory analyses are presented in Appendix F and relevant data is detailed in the discussion.

At boreholes where samples were taken, hydraulic head was measured using an electric water level tape. Samples were taken after purging, except in the case of active production wells, where purging was not necessary. Stagnant water in the well column was purged either manually, using a PVC bailer, or by a micropurge pump. Wells were purged until the key parameters of EC ($\mu S/cm$), pH, temperature (°C) and dissolved

Table 6.1. Number of field sampling rounds and the number of samples taken for each analyte. The 2006 samples were taken by Greg Jones of the South Australian Department of Water Land and Biodiversity Conservation.

Year	Month	Major ions	¹⁸ O & ² H	13 C & 14 C	87 Sr	36 Cl	
2004	Nov.	10	10				
2005	Oct. & Nov.	43	35				
2006	Feb. & June	34	34				
2007	April & May		34	13	34		
2008	Feb.			2		9	

oxygen (mg/L), had stabilised. These parameters were measured in a flow cell using an Orion multimeter. The EC probe was frequently calibrated against standards of values $1412\mu S/cm$ and $12880\mu S/cm$ and measurements were made standardised to 20°C. The pH probe was frequently calibrated against buffers 4, 7 and 10 and the dissolved oxygen probe was automatically calibrated in air.

Samples were tested for several parameters in the field in order to minimise the time in which chemical reactions could alter the composition of the water. Dissolved inorganic carbon (DIC) concentration in the waters was measured in the field using HACH digital titration kit. The speciation of carbon is related to the pH of the solution (see Figure 6.1). Bicarbonate dominates in the pH range from 7 to 10 and carbonate dominates when pH is 11 and above. In this study, the total alkalinity, being the sum of the concentration of HCO_3^- and CO_3^{2-} , was measured in the field by titrating the samples with sulfuric acid to a 4.5 pH endpoint. The third carbon species, carbonic acid, which dominates at the lower pH range, was not directly measured, however, it is related to the concentration of dissolved CO_2 . Dissolution of CO_2 into water is proportional to the partial pressure of CO_2 in the gas phase in equilibrium with the water. Once dissolved, $CO_{2(aq)}$ reacts with the water to form carbonic acid, H_2CO_3 , which can eventually dissociate to HCO_3^- , although both are slow processes. Dissolved CO_2 was measured in the field by titrating samples with sodium hydroxide to a 8.3 pH endpoint.

High-density polyethylene (HDPE) bottles (125ml) were used to collect samples for laboratory analysis of cations, anions, ¹⁸O, ²H, ¹³C and ⁸⁷Sr. Bottles were filled to the exclusion of air bubbles and caps were taped to minimise atmospheric contamination and degassing. Cation and ⁸⁷Sr samples were filtered with 0.45μ m cellulose nitrate filters. Cation samples were acidified with ultrapure 16N nitric acid (HNO₃) in the field before laboratory analysis to prevent loss through sorption.

Cations were analysed using an Inductively Coupled Plasma Mass Spectrometer (ICP-MS) at the Australian National University. Anions were analysed at Monash University



Figure 6.1. Carbon species activity with varying pH range, after Drever, 1997).

using a Metrohm Ion Exchange Chromatograph (IC). Cation and anion concentrations were reported as mass-volume ratio (mg/L). Concentrations were converted to mmol/L for molar ratio calculations and meq/L for charge balance error calculations.

The charge balance errors for almost half the samples (41 samples of 85) were within acceptable limits, falling between -10% and +10%. There is only one sample which has a charge balance error less than -10%, however, the remaining 51% of the samples produced charge balance errors greater than 10%. This implies that an important anion is missing from the analyses, or an anion has been under-represented in terms of the concentration measured. There are two likely candidates, PO_4^{3-} and total alkalinity ($HCO_3^-+CO_3^{2-}$). Only 7 of the 85 samples showed a peak correlating to the presence of the phosphate ion on the ion chromatograph and one of these has a negative charge error balance, hence it is unlikely that phosphate is the source of the imbalance between anions and cations in most of the solutions.

The more likely explanation is that the actual value of total alkalinity is higher than that measured by the titrations in this study. The worst of the high charge balances, meaning those above +15%, are all very fresh waters; Cl⁻ concentrations are under 36 meq/L. The lowest concentration within this group is 2.45 meq/L, where a difference of just 0.3 meq/L would produce the unacceptable error of 12%. If this entire imbalance were due to alkalinity, it would correspond to 10.5 mg/L as HCO₃⁻. Given that the precision



Figure 6.2. The variability with pH of relative proportions of carbon species measured in field titrations, plotted as points. The expected proportions for a closed system are plotted as curves.

for the alkalinity field titration is approximately 5 mg/L as HCO_3^- , it is likely that error in the titration is responsible for at least a component of the change imbalance in most samples.

Another possibility is that addition of CO_2 has affected the measurement of alkalinity. Figure 6.2 shows the molar proportion of the measured carbon species: total alkalinity and dissolved CO_2 . It can be seen that, for many samples, the concentration of CO_2 is higher than is stable at the given pH. The pH was taken at the time of sampling and therefore represents the conditions in the aquifer. In some cases, HCO_3 and CO_2 were measured by titration some time after collection, due to constraints on field resources. It may be the case that some CO_2 bubbles were suspended at the time of sampling, and therefore did not contribute acid to the pH measurement, but did eventually dissolve into the solution as CO_2 or H_2CO_3 . Hence, by the time titrations were conducted, the CO_2 and H_2CO_3 proportions represent a lower pH. Normally, given the buffered conditions in the aquifer, this acidity would have dissociated to HCO_3 , contributing to a greater concentration of anions.

For the sake of clarifying the complications just described regarding alkalinity and carbon speciation, the total alkalinity was recalculated according to the charge balance. The pH value was then used to calculate the expected proportions of carbon species. The recalculated values compare much more favourably with the public data than the original titrations. For example, the ratios of sodium to total alkalinity of the two data sets are similar (see Figure 6.3). All data, calculation results and titration data, are presented in Appendix F.

The molar ratios of ⁸⁷Sr to ⁸⁶Sr isotopes of the strontium ions in solution, expressed



Figure 6.3. The molar ratio between sodium and total alkalinity plotted against chloride for groundwater from all the aquifers of the basin. The titration- and calculation-derived values from this study are compared to the state data set.

as 87 Sr/ 86 Sr, were measured at La Trobe University using a solid source mass spectrometer. The stable isotopes, 18 O and 2 H, of the water particles and the 13 C composition of the DIC were analysed using a dual inlet gas-source mass spectrometer. The oxygen and hydrogen of the water samples were allowed to equilibrate with CO₂ and hydrogen gases respectively. The method followed for oxygen analysis is that of Epstein and Mayeda (1953). DIC was extracted as CO₂ via acidification under vacuum. The stable isotopes were measured against in-house standards with known compositions comparative to IAEA standards. Corrections were performed on one dataset due to the difference in the standards used (these corrections are detailed in Appendix L). The results were reported as δ^{18} O-VSMOW, δ^{2} H-VSMOW, and δ^{13} C-VPDB, with respective analytical precision being: $\pm 0.15\%$, $\pm 1\%$ and $\pm 0.2\%$.

Measurement of ¹⁴C composition of the DIC in water samples was done using one of two methods, accelerator mass spectrometry (AMS) and direct absorption (DA). All analyses were conducted by CSIRO Land and Water or their partners (for example, two samples were analysed using AMS at the Australian National University). Samples with adequate concentrations of DIC (HCO₃ in excess of 200 mg/L) were measured using DA, which generally required 40 L of water, collected in two 20 L HDPE carbouys. Leaney (1994) detailed the extraction and liquid scintillation counting method for direct absorption. Waters with low concentrations of DIC were analysed using AMS following Davie *et al.* (1989a); samples were collected in 1L HDPE bottles. Results were reported as percent modern carbon (%mC or pmC), having been normalised to the δ^{13} C value of the oxalic acid standard. Error ranged from ±0.1 to ±1.4%mC. A value of 100% modern carbon is defined as 13.65 dpm/g (disintegrations per minute per gram) and represents the activity of ¹⁴C in vegetation growing around 1890 (Clark & Fritz, 1997). For groundwater dating purposes, 100%mC represents the initial activity of ¹⁴C in the atmosphere before the bomb-related pulse of elevated activity after 1950. This initial activity, A₀, is used to calculate the raw radiocarbon age, using Equation 6.1.

$$t = \frac{\ln(\frac{A_t}{A_0})}{-\lambda} \tag{6.1}$$

where t is the time or age (radiocarbon years), A_t is the measured activity of the sample (%mC) and λ is the decay constant for ¹⁴C, which equals $1.209 \times 10^{-4} yr^{-1}$. The %mC results were converted to ages in this way and were then corrected for "dead" carbon contamination using the δ^{13} C values of the samples (discussed in Chapter 7).

Groundwater samples for ³⁶Cl were collected, unfiltered, in acid-washed, 1L HDPE bottles. Analysis was performed at the ANU using the 14UD tandem electrostatic ion accelerator. Preparation of the sample involved precipitation of the dissolved Cl- ions as silver chloride. Sulphur, having a similar atomic mass to chlorine, is a conflicting element in the analysis and was removed by precipitation as barium sulphate. The complete analytical methodology used was that of Fifield *et al.* (1987). The results were reported as a ratio of ³⁶Cl atoms to total chlorine atoms, denoted as: ³⁶Cl/Cl×10⁻¹⁵. Error ranged from $\pm 1.6 \times 10^{-15}$ to $\pm 2.4 \times 10^{-15}$.

6.3 Salinity and hydrogeochemistry

In the sections below the major ion concentration and electrical conductivity data from the public records and from this study are used to characterise each of the major aquifers. The geochemistry of the water is compared to the overall salinity by presenting the electrical conductivity (EC) maps in these sections. The use of EC as a proxy for salinity was influenced by the abundance of conductivity records in comparison to salinity records. The influence of temperature on conductivity was overcome by presenting all the EC data as normalised to 20°C. Hence, the EC values of groundwater samples presented in this Chapter are an excellent tool for assessment of the salinity of the water. However, as discussed in Chapter 4, the conversion of EC to salinity is not a simple process and



Figure 6.4. Cl⁻/Br⁻ ratio vs Na⁺/Cl⁻ ratio of groundwater samples from this study.

the geochemical composition of the dissolved constituents can influence the relationship. For this reason, the direct comparison of specific salinity values and EC values in this discussion is avoided where possible.

The comparative concentration of major ions are used to assess how the groundwater chemistry changes temporally and spatially. For this analysis to be meaningful, some indication of the conservative nature of the chloride ion should be given as it is usually assumed to be conservative in such discussions. To assess this, the trends in the Cl^-/Br^- ratio and the Na⁺/Cl⁻ ratio of the groundwater are introduced briefly here.

The molar Cl^-/Br^- ratio of seawater is 655 ±4 (Alcalá & Custodio, 2008). Rainfall at the coast usually has a similar ratio to this and will have increasingly lower ratios inland, which can be lower than 360 (Davis *et al.*, 1998b). Despite this, rainfall ratios as high as 960 have been measured in Victoria (Bormann, 2004) and the ratio of water from the estuary of the Hopkins River, obviously closely reflecting both rainwater and seawater, measured in this study is approximately 720. The elevated local ratios are apparent throughout the groundwater of the Otway Basin (Figure 6.4), although some samples from inland areas have lower ratios. There are numerous processes of the atmosphere or unsaturated zone that may be responsible for this regional fractionation of the Cl^{-}/Br^{-} ratio in the meteoric source water (Fabryka-Martin, 2005). For example, the enrichment of Cl⁻ in most of the groundwater samples when compared to the established marine ratio may be due to a regional effect, such as the addition of halite in atmospheric dust. Additionally, although a rare process, sorption of bromide on organic matter in the soil zone linked to land clearing could also be influential (Bennetts, 2005; Alcalá & Custodio, 2008). These processes are distinct from significant dissolution of geological deposits of halite along the flow path, which results in even higher molar Cl^{-}/Br^{-} ratios of at least 2200 (Davis et al., 1998b; Bear et al., 1999). Two samples in Figure 6.4 have such high Cl⁻/Br⁻ ratios, indicating that halite dissolution is important in their evolution. They are from boreholes screened in the shallow aquifers and therefore halite is likely to have been in the unsaturated zone. They have low salinity (chloride concentration < 100 mg/L) and so halite dissolution has not been extensive. Overall, the Cl⁻/Br⁻ ratios of the groundwater indicate that the majority of the chloride dissolved was originally introduced by wet or dry precipitation (i.e. rain or dust). This has the further benefit of confirming Cl^{-} as a relatively conservative solute in the groundwater system, to which the behavior of other ions may be compared. This conclusion is also supported by similar analysis of Cl⁻/Br⁻ ratios in the Corangamite region (Cox et al., 2007).

The one of the most useful ions in this sense is sodium. Molar Na^+/Cl^- ratios of groundwater may be similar to the marine ratio (0.86; Bear *et al.*, 1999), reflecting a composition dominated by rainfall-derived cyclic salts; may be closer to 1, indicating a dominance of halite dissolution; or may rise or fall according to addition or loss of sodium independent of chloride, indicating water-rock interaction. Such water-rock interaction has been observed in other hydrogeological studies of the Otway Basin (e.g. Johns, 1968; Blake, 1980) and can include reactions such as silicate dissolution, clay mineral formation and ion exchange. Figure 6.5 shows the changes to the Na^+/Cl^- ratio of the groundwater of the Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers with changes in salinity and relative distance along the flow path (records from Victorian section only). This diagram provides a preliminary look at the chemistry of the water of these aquifers and the variation in the ionic ratio confirms that water-rock interaction processes have affected the groundwater. Specific processes and the sources of the solutes are discussed in further detail below where the aquifers are addressed each in turn.



Figure 6.5. Na^+/Cl^- ratio of groundwater samples from the Upper Tertiary Carbonate Aquifer and the Lower Tertiary Sandy Aquifer in the Victorian records compared to changes in chloride concentration and relative distance along the flow path.

6.3.1 Salinity distribution in Surficial Aquifers

Rather than one hydrostratigraphic unit, this section deals with salinity of the groundwater of all the surficial aquifers in the basin. Similar to the watertable map presented in Chapter 4, the map of Figure 6.6 represents the EC of groundwater samples from any wells in the basin screened in the uppermost 20 m of the subsurface. These bores are screened in almost all the hydrostratigraphic units of the basin and therefore this map can be used to delineate the effect of lithofacies on salinity distribution. Especially featured in this Section are the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers because of their dominance of the surficial geology (refer to Figure 2.1). Additionally, because this map depicts all the EC variation of groundwater close to the surface, it provides insight into some processes of groundwater-surface water interaction.

The shallow groundwater of the Otway Basin has a highly variable salt concentration, as indicated by the EC values in Figure 6.6 which range from a few hundred μ S/cm to more than 40 mS/cm. The lower end of this range is similar to typical EC values of rain water and the upper limit is close to those of seawater, which is on average approximately 50 mS/cm.

It can be seen from Figure 6.6 that the Gambier Sub-basin hosts the majority of the fresh shallow groundwater in the basin. The surficial geology of this region is dominated by the dune and swale deposits of the Plio-Pleistocene Heterogeneous Aquifer which are oriented sub-parallel to the coast (refer to Figure 2.1). The Upper Tertiary Carbonate Aquifer is also exposed in much of the south of the Sub-basin and in the north at the upthrown block of the Kanawinka Fault. The Pliocene-Quaternary Volcanic Aquifer is restricted to small outcrops at Mt Burr, Mt Schank and Mt Gambier. The overall low salinity of the Gambier Sub-basin shallow groundwater is a product of fast deep drainage of precipitation. The mechanisms of fast effective recharge in this area have been discussed in detail in Chapter 4, especially regarding Sub-aquifer One of the Plio-Pleistocene Heterogeneous Aquifer and the Upper Tertiary Carbonate Aquifer. Importantly, the secondary karstic porosity of the region plays a role in promoting vertical drainage. This prevents the prolonged exposure of water in the soil zone where its solutes may be concentrated by evapotranspiration. The idea that localised recharge occurs at intervals all along the regional coastward flow system is supported by the persistence of fresh groundwater throughout the Gambier Sub-basin. In other words, because the map in Figure 6.6 is not depicting a regional flow system, the classic case of accumulation of dissolved solids along the flow path (e.g. East Midlands Triassic aquifer, UK) is not observed.

Some small areas of more saline shallow groundwater exist in the Gambier Sub-basin; they are associated with the low-lying inter-dune swale deposits. The swales are barriers to drainage because they are oriented orthogonal to regional coastward flow and are composed of the low-permeability sediment of Sub-aquifer Two of the Plio-Pleistocene Heterogeneous Aquifer. The swales are often occupied by lagoons which are localised discharge zones. Because the local vertical hydraulic gradients are upwards at these locations, the lagoon water and the shallow groundwater are affected by evaporative concentration. On the other hand, some ephemeral lagoons can develop as localised recharge zones in response to rain events or wet periods and are often not significantly concentrated by evapotranspiration because deep drainage from them occurs relatively quickly.

This last mechanism is important in maintaining the high quality of the shallow groundwater and in turn, the quality of the groundwater in regional flow systems to which the





surficial system is connected. One implication of this observation is that future climate change may pose a threat to the quality of the groundwater resource in the long term. Currently, evaporation rates are greater in the hot, dry summer and recharge usually occurs in the more humid winter (Chapter 3). However, if winter rainfall is to become less reliable and both summer and winter temperatures are to increase (IPCC, 2007), an increased rate of evaporative concentration of dissolved solids may also be expected.

The upper Wannon region, north of the Tyrendarra Sub-basin, is mostly blanketed by the Pliocene-Quaternary Volcanic Aquifer and contains shallow groundwater that is generally brackish in areas of the first phase volcanics and fresh in the regions of the second phase volcanics. The latter are most notably represented in this region by the deposits associated with Mt Rouse and Mt Napier, whose eruption centres lie on the groundwater divide with the sub-basins to the south. This relationship with salinity is due to the greater degree of weathering generally present in the older volcanic rocks and therefore a greater development of clay minerals; although the younger basalt can also be deeply weathered (Joyce, 1988). Active weathering has been demonstrated by hydrogeochemistry work on the Pliocene-Quaternary Volcanic Aquifer in other regions and clay minerals include kaolinite, illite and smectite (Bennetts et al., 2006). The relationship between the phases of volcanism and the groundwater salinity has been recognised in the past (Gill, 1989) although sometimes attributed to ion filtration by the clay minerals (Maddocks, 1967b). While ion exchange on clay minerals is an important mechanism controlling groundwater chemistry in the Otway Basin (Johns, 1968; Blake, 1980), it is not the cause of salinity accumulation. Rather, the poor hydraulic conductivity of the aquifer where it is enriched with clay (Thompson, 1972) inhibits recharge and allows evapotranspirative concentration of the groundwater solutes which are derived from precipitation, dust and water-rock interaction.

Bennetts *et al.* (2006) observed the same clay-related process of salt concentration in groundwater of the Pliocene-Quaternary Volcanic Aquifer and fractured basement rocks to the northeast of the upper Wannon region, near Glenthompson (refer to Figure 4.5). Although technically outside the boundaries of the Otway Basin because of the lack of Tertiary strata under the Quaternary aquifers, some shallow groundwater in that area flows southwestward into the upper Wannon region through the volcanic rocks. Therefore, some import of solutes is to be expected. In the Glenthompson region, the local- or intermediatescale discharge zone, Cockajemmy Lakes, is characterised by saline water of primary origin which has developed over the past 8000 years at least (Bennetts *et al.*, 2006).

Similar mechanisms of salinity development are responsible for the high EC values of shallow groundwater in the northern areas of the Tyrendarra, Port Campbell and Barwon Downs Sub-basins and for those in the Corangamite region. One notable example is a zone of groundwater with EC largely greater than 7000 μ S/cm in the mid-reaches of the Hopkins River and Mt Emu Creek. This area is mostly occupied by basalt flows, with some of the more dilute water correlating to outcrop of the Hanson Plain Sand of the Plio-Pleistocene Heterogeneous Aquifer. This zone of high EC is punctuated in its centre by the fresh groundwater hosted in the scoria and stony rises of Mortlake (refer to Figure 4.5) and is surrounded in the south and east by fresh groundwater derived from other topographic highs, e.g. Mt Noorat near Lake Keilambete. Therefore, the relationship of salinised groundwater is not just with the lithofacies of the volcanic rocks, but also with the topographic highs of certain eruption centres. This highlights the importance of the local scale flowsystems discussed in Chapter 4 that are driven by topographic relationships where small scoria cones are greatly differentiated from the surrounding low lying plains. When comparing Figure 6.6 to the topographic features in Figure 1.2, it is clear that the low relief of this region prevents effective development of runoff and drainage and promotes the evapotranspiration of shallow groudnwater, concentrating solutes.

In contrast, the southern regions of the Tyrendarra and Port Campbell Sub-basins, as well as the Geillbrand River region, contain shallow groundwater that is mostly fresh. These areas are generally characterised by thin surficial cover over the Upper Tertiary Carbonate Aquifer or, in the case of the Gellibrand River region, outcrop of the older Aquifers. The relatively high hydraulic conductivity of these units allows for efficient deep drainage and prevention of excessive solute concentration at the surface. There are two notable exceptions within these regions. The first is groundwater of EC up to 5,000 μ S/cm hosted partly within the exposure of the Upper Tertiary Aquitard at the watertable divide on the western edge of the Gellibrand River region. The second is a wedge of saline groundwater (40,000 μ S/cm) close to the concentration of seawater (50,000 μ S/cm) at Warrnambool, near the mouth of the Hopkins River. The water is likely to be a mixture of fresh groundwater and seawater, some of which may have circulated through the saline estuary of the Hopkins River (EC > 6,000 μ S/cm; refer to Section 4.6).

Apart from Warrnambool, the regions with the most saline shallow groundwater are the aforementioned Corangamite region and Barwon Downs Sub-basin. The brackish water accumulates in the catchment of the Corangamite region via the mechanisms previously discussed and groundwater discharge of the closed system at Lake Corangamite results in hypersalinity (EC > 200,000 μ S/cm). However, the lake was up to ten times more dilute in the 1960s, before the diversion of the Woady Yaloak River (Timms, 2005). These areas are prone to salinity concentration at the surface by evapotranspiration because of the poor drainage network, the low relief topography, and the high clay content of the soil in some areas.

6.3.2 Geochemistry of the Pliocene-Quaternary Volcanic Aquifer

There are two major sources of solutes for the groundwater of the Pliocene-Quaternary Volcanic Aquifer: cyclic salt, including dry deposition of dust and solutes in rain, and waterrock interaction. Both sources are characterised below and compared to the groundwater chemistry observed in the aquifer. The water-rock interaction of the Pliocene-Quaternary Volcanic Aquifer is especially described in detail for the reason that the widespread surficial cover of this aquifer results in its influence on the composition of groundwater entering the underlying units.

The major ion chemical composition of the groundwater of the Pliocene-Quaternary Volcanic Aquifer is presented in Figure 6.7. Anions are dominated by chloride, although more than 50% bicarbonate is common. Cations are variable; most samples are dominated by sodium, but some have proportions of magnesium and calcium of up to 60% and 65% respectively. The high-salinity water is Na⁺Cl⁻ or Na⁺Mg²⁺Cl⁻ type, whereas the fresh water is $HCO_3^-Ca^{2+}$ type.

When compared to the composition of precipitation in the coastal and inland regions of the basin (Hutton & Leslie, 1958; Figure 6.7), the groundwater commonly contains relatively more chloride or more bicarbonate, and in some instances is enriched in sulphate. The cations of the groundwater are each enriched relative to their proportions in precipitation in different samples. Evapotranspiration can concentrate the solutes in groundwater derived from rainfall and dry deposition. This process acting alone produces water with a very similar composition to precipitation but a greater concentration. There are a number of samples that fit this criteria. However, the differences evident in Figure 6.7 between groundwater and source meteoric water indicate that water-rock interaction is also an important control on the chemistry.

Most of the volcanic rocks of the basin, be they in the form of lava flows, scoria cones or stony rises, are basaltic. The basalt of the Pliocene-Quaternary Volcanic Aquifer typically contains phenocrysts of olivine, augite and alkali feldspar (e.g. sanidine and anorthoclase), and has a groundmass composed of silica (glass), augite, plagioclase (labradorite) and other feldspar (Joyce, 1988). Weathering products observed include iron oxide, montorillonite (of the smectite group) and chlorite (Joyce, 1975; 1988); and those expected from hydrogeochemical calculations are kaolinite, illite and smectite (Bennetts *et al.*, 2006). The other main lithology of the aquifer is the tuff of its maars. The weathering of ash can involve many of the reactions listed below concerning weathering of basalt, including dissolution of glass and degradation of feldspar. Country rock components of the tuff are usually dominated by the Tertiary limestone and marl and therefore the dissolution of carbonates, i.e. calcite and dolomite, is expected to be a common reaction (Equations 6.6



Figure 6.7. Ternary diagrams showing the relative abundance of the major ions for all Pliocene-Quaternary Volcanic Aquifer records in Victoria. The relative size of the points relates to the total concentration of ions (meq/L). Precipitation composition (including dry deposition) is represented by two samples from Hutton and Leslie (1958; size of symbols not to scale).

& 6.7).

Of the main constituents of basalt, being olivine, pyroxene and feldspar, olivine is the most unstable, and therefore most soluble, and feldspar is the least soluble (Kehew, 2001). The relative order of solubility within feldspars from most soluble to least soluble is anorthite, albite and then orthoclase (Krauskopf, 1979). The specific reactions for weathering of basalt minerals are detailed below, followed by a discussion of their relevance to the groundwater composition in the Pliocene-Quaternary Volcanic Aquifer.

Volcanic glass, obsidian, does not contain minerals as it lacks crystal structure. Its elemental composition reflects that of other felsic igneous rocks such as rhyolite or granite and is generally dominated by silica (SiO₂). Amorphous silica such as this is more soluble than quartz. The common maximum dissolved concentration of silica in natural waters at the earth's surface is 60 mg/L (Bricker & Garrels, 1967), although the theoretical solubility of amorphous silica at 25° C is 115 mg/L (Hem, 1985). As the silicic acid concentration of the groundwater increases, the volcanic glass becomes less soluble.

Olivine can weather completely via congruent dissolution to silicic acid and magnesium/iron ions with the aid of carbonic acid in the groundwater (Andrews *et al.*, 2004), as can be seen in the case of forsterite:

$$Mg_2SiO_4 + 4H_2CO_3 = 2Mg^{2+} + 4HCO_3^- + H_4SiO_4$$
(6.2)

Although, the Fe^{2+} in olivine can be oxidised to Fe^{3+} in the right conditions and result in production of a solid phase, e.g. goethite:

$$MgFeSiO_4 + 3H_2O + CO_2 = FeO(OH) + Mg^{2+} + H_4SiO_4 + HCO_3^{-}$$
(6.3)

thus, olivine weathering consumes carbonic acid and contributes magnesium, silicon, bicarbonate and some iron to solution.

The chlorite present in the volcanic rocks is an intermediate product of pyroxene weathering, ultimately the chlorite weathers to other clay minerals: kaolinite, smectite or illite. The latter process releases cations Mg^{2+} and Fe^{2+} into solution from the hydroxides between chlorite's silicate sheets (Sumner, 2000), or can also oxidise iron to form goethite (Aspandiar & Eggleton 2002).

The incongruent dissolution of feldspar minerals takes metals from the mineral structure and ionises them in solution, leaving the mineral altered, usually to clay. One example is albite weathering by water and carbonic acid, producing kaolinite (solid) and sodium and bicarbonate ions and silicic acid in solution, as described by the equation:

$$2NaAlSi_{3}O_{8} + 2H_{2}CO_{3} + 9H_{2}O = Al_{2}Si_{2}O_{5}(OH)_{4} + 2Na^{+} + 2HCO_{3}^{-} + 4H_{4}SiO_{4} \quad (6.4)$$

Similar reactions exist for other feldspar minerals by substituting the metals (sodium above) on both sides of the equation. Alternatively, feldspar weathering may result in the formation of other clays, e.g. smectite from plagioclase. In another variation of the weathering path, montmorillonite, part of the smectite group, can form from kaolinite (Blake, 1980) thus:

$$3.5Al_2Si_2O_5(OH)_4 + Na^+ + 4H_4SiO_4 = 3Na_{0.33}Al_{2.33}Si_{3.67}O_{10}(OH)_2 + H^+ + 11.5H_2O$$
(6.5)

This reaction decreases groundwater pH while consuming sodium and silicon in solution. As with Equation 6.4, the sodium above may be replaced by magnesium or calcium to describe other clay reactions.

The formation of the clay assemblage is a product of the parent rock, as discussed above, but is also subject to the leaching conditions and the amount of biologicallymediated weathering that occurs. In the reactions listed so far, acid hydrolysis is facilitated by carbonic acid, which is produced in the groundwater by dissolution of CO_2 in the soil zone during recharge. However, also during recharge, organic acids, such as fulvic or humic acid, are dissolved in the groundwater and produce ligands which can form complexes with the cations of silicate minerals (Kehew, 2001). In this way, otherwise insoluble ions, such as aluminium, can be released into solution. Further, different clay minerals are predisposed to form in certain drainage environments according to the silica:aluminium ratio in their molecular structure (Andrews *et al.*, 2004). For example, formation of kaolinite usually requires some export of dissolved silica via leaching to reduce the Si:Al ratio to 1:1, whereas smectite, whose ratio is closer to 2:1, often forms in poorly drained areas.

Using the general outcomes of the mineral reactions described above, the changes in the geochemistry of the groundwater of the Pliocene-Quaternary Volcanic Aquifer shown in Figure 6.7 can be interpreted. The evolution of the groundwater is described with the assumption that concentration of solutes occurs concurrently with water-rock interaction. Therefore, the fresh water samples are seen to represent the early stages of the evolution and the most saline the final stages.

The freshest groundwater in the aquifer usually has a greater proportion of calcium and bicarbonate than cyclic salt inputs (Figure 6.7). This is the result of acid hydrolysis reactions that are faster than others, namely: incongruent dissolution of anorthite which is more soluble than other feldspars (Equation 6.4 adjusted for Ca^{2+}); weathering of Cabearing pyroxene; and dissolution of soil carbonates and calcite or dolomite from tuff (Equations 6.6 and 6.7).

Further along the evolution of the groundwater, the addition of sodium and magnesium by olivine and pyroxene dissolution and by feldspar weathering is important. These reactions also produce bicarbonate ions. However, concurrent addition of cyclic salt and evaporitic concentration renders the overall groundwater composition more similar to precipitation, i.e. containing a greater proportion of chloride.

The precipitation of calcite once the water reaches saturation may also have removed calcium and bicarbonate from solution at this stage. Calcite and dolomite are precipitating in western Victorian lakes (Maddocks, 1967b) and in lakes of the Coorong region (Von Der Borch, 1965) in South Australia. During field work for this thesis, modern biogenic calcrete, formed by a process similar to that of stromatolites, was observed at the lake shore and bed of Lake Bullen Merri. Calcium could also be lost from solution via cation exchange as it has greater sorption affinity than Mg^{2+} or Na^+ .

The final stage of evolution involves further silicate weathering and calcite precipitation as well as cyclic salt input from ongoing recharge and concentration via evapotranspiration. Additionally, cation exchange and clay alteration are the most likely factors controlling whether the groundwater is dominated by magnesium or sodium. The bulk composition of the basalt of the Pliocene-Quaternary Volcanic Aquifer contains up to 2 times more magnesium than sodium (by weight, Price *et al.*, 1997). Therefore, ongoing weathering might be expected to produce a magnesium-type water. However, cation exchange could favour a sodium-type water as magnesium has a greater affinity for sorption under normal conditions. In high salinity water, sodium has a greater propensity for sorption, favouring a magnesium-type water.

Furthermore, removal of either ion could eventuate through the conversion of kaolinite to montmorillonite (e.g. Equation 6.5). Kaolinite has been observed in areas of higher topographic elevation in the basalt plains of the Tyrendarra Sub-basin because those areas are relatively well drained (Gibbons & Gill, 1964). Montmorillonite has been confirmed in other regions (Joyce, 1975) and because they are poorly drained, these regions are also where more solutes are concentrated in the water.

The precipitation or dissolution of halite can affect the sodium and chloride proportions in the groundwater. However, from the preliminary investigations into $Cl^-/Br^$ ratios above, it is apparent that the process of dissolution is not widespread. Hypersaline sodium-chloride water, e.g. that in Lake Corangamite, is in equilibrium with halite and precipitation is widespread in the lake environment. However, this does not act to reduce the net sodium or chloride in solution because of re-dissolution. The silicon/chlorine ratio of the groundwater can be used to further confirm the important role silicate weathering plays in influencing the chemistry of the groundwater in the Pliocene-Quaternary Volcanic Aquifer. The ratio in rain is dictated by the amount of continental dust containing silica in the atmosphere. This amount varies greatly, but some measured values for coastal or natural ratios range from 0.02 to 0.04 (Gibbs, 1970; Probert, 1976) and some values for dust-contaminated precipitation are 0.06–0.28 (Pelig-Ba *et al.*, 2001). Four groundwater samples from the volcanics were analysed for silicon concentration and their Si/Cl ratios were in the range 0.07–0.78. Assuming that chloride is conservative, these ratios indicate that silicic acid has been added to solution in the groundwater by all or any of the water-rock interactions listed above (except Equation 6.5).

Not explained by any aspect of this water evolution are the changes to the sulphate abundance in the groundwater relative to the source precipitation (Figure 6.7). A few samples from the Pliocene-Quaternary Volcanic Aquifer show dominance of sulphate over other anions and their sulphate-chloride ratio is significantly higher (up to 0.6) than that of coastal rainfall or marine water (0.056). Therefore, a source of sulphur other than cyclic salts is necessary. It is well accepted that relative SO_4^{2-} concentrations in rainfall increase inland, as represented the increase of SO_4^{2-}/Cl^- molar ratio with continentality (Junge & Werby, 1958) and, in Australia, with aridity (Herczeg *et al.* 2001). This is due to the increasing component of continental dust. Continental dust is discounted as a source here because its contribution is included in the precipitation compositions plotted in Figure 6.7. Another alternative source is weathering, however, gypsum is known to be rare in the basin and is notably absent from the volcanic lakes (Maddocks, 1967b).

The source of the sulphur is most likely volcanic H_2S emanations from depth, delivered via faults and joints, or H_2S produced by biological reactions within the aquifer. Most of the groundwater samples lie between pH 7 and 9. At this pH range, H_2S is not stable. In oxidising conditions, as would be expected in this unconfined aquifer, H_2S oxidises to SO_4^{2-} .

Nitrogen concentration in the groundwater (primarily as nitrate) was not analysed in this study but is briefly discussed in Section 6.4.1.

With the exception of sulphur and nitrogen, the solutes of the groundwater of the Pliocene-Quaternary Volcanic Aquifer have been found to be sourced from cyclic salt deposition (wet and dry) and from water-rock interaction enabled by both inorganic acid hydrolysis and biologically-mediated reactions. As the volcanic lakes are linked to the groundwater flow system (Chapter 4), these sources apply also to them. Cyclic salt as a major source of solutes in the lakes is supported over a connate origin by several previous authors (e.g. Anderson, 1941; Maddocks, 1967b).

6.3.3 Upper Tertiary Carbonate Aquifer

The Upper Tertiary Carbonate Aquifer hosts large volumes of fresh water, particularly in the Gambier Sub-basin (Figure 6.8). There are, however, some regions where the EC is greater than 3,000 μ S/cm, indicating poorer quality water. In general these regions coincide geographically with overlying brackish shallow groundwater (Figure 6.6). This pattern of groundwater EC in the Upper Tertiary Carbonate Aquifer being similar to that of the watertable indicates that the flow systems of the aquifer and the overlying units are integrated. The fact that an increase in salinity along the regional flow path is not observed reveals that the local- and intermediate-scale flow systems are important in controlling the groundwater quality.

Also visible in Figure 6.8 are three small zones at the coast with very high salinity, i.e. $EC = 30,000 \ \mu S/cm$ or higher, which are due to seawater intrusion, or the presence of the seawater interface onshore. They are at Port Macdonnell, Port Fairy and Warrnambool. Such high salinity groundwater was detected at shallower depths only at the Warrnambool site (Figure 6.6). Refer to Section 4.6 for more on the interface location.

The major ion chemistry of the groundwater of the Upper Tertiary Carbonate Aquifer is presented in Figure 6.9, along with two points representative of inland and coastal rainfall in the region (Hutton & Leslie, 1958). On the whole, the saline samples of groundwater are Na⁺Cl⁻ type and can be enriched in these ions relative to cyclic salt proportions. Moderate salinity groundwater is similar in composition to rainwater. Low salinity groundwater tends to be $Ca^{2+}HCO_3^-$ type and is very enriched in these ions with respect to the composition of rainwater. Sulphur is the weakest anion and rarely constitutes greater than 10% of the total. Conversely, the least dominant cation, magnesium, usually has a concentration greater than 10%.

Three cross sections, one in each of the major sub-basins, compare the EC of the groundwater to its typical chemical composition (Figures 6.10, 6.11 and 6.12). However, a significant amount of variation from the representative compositions occurs in the Upper Tertiary Carbonate Aquifer. This is in part due to the differing groundwater compositions that result from diffuse recharge and from point source recharge through karst features. Leaney and Herczeg (1995) and Herczeg *et al* (1997) showed that variation in chemical properties of groundwater of the Upper Tertiary Carbonate Aquifer correlates to proximity to karst features, such as sink holes. Low but variable chlorinity and CO₂, a low Mg^{2+}/Ca^{2+} ratio and a negative $SI_{calcite}$ value are characteristics of groundwater near point source recharge sites. Conversely, groundwater that has recharged through the soil profile has higher and mostly stable chlorinity due to mixing, higher CO₂ from microbial and root respiration in the soil zone, a $SI_{calcite}$ close to zero and a higher Mg^{2+}/Ca^{2+} ratio







Figure 6.9. Ternary diagrams showing the relative abundance of the major ions for all Upper Tertiary Carbonate Aquifer records in Victoria. The relative size of the points relates to the total concentration of ions (meq/L). Precipitation composition (including dry deposition) is represented by two samples from Hutton and Leslie (1958; size of symbols not to scale).

because calcite precipitation.

Port Campbell Sub-basin

The hydrogeochemistry cross section for the Port Campbell Sub-basin is presented in Figure 6.10. This diagram also contains charts of chemical compositions for two precipitation samples representative of inland and coastal rainfall in the Victorian Section of the Otway Basin (Derrinallum and Warrnambool; Hutton & Leslie, 1958). The groundwater of the Upper Tertiary Carbonate Aquifer undergoes several changes in salinity along the regional flow path, as can be seen from the EC contours, which are due to the processes discussed above involving local flow systems and accumulation of salt in low lying or clay rich zones.

The groundwater composition at location **a** (Figure 6.10) is similar to inland rainfall with the difference of a greater proportion of magnesium and calcium. These solutes are derived from the weathering of the overlying basalt, through which this groundwater has recharged. A smaller proportion of bicarbonate here compared to the inland rainwater is interesting as bicarbonate is a product of the weathering reactions that have delivered the Mg^{2+} and Ca^{2+} . If chloride is conservative, this change signals removal of bicarbonate, possibly by calcite precipitation as intergranular cement in the aquifer.

Further along the flow path at \mathbf{b} , the groundwater composition relates to the physical mixture the water consists of, namely a combination of water that has flowed from \mathbf{a} with other water more recently recharged through the Pliocene-Quaternary Volcanic Aquifer. Flow through the limestone of the aquifer results in carbonate dissolution, most likely calcite dissolution:

$$CaCO_3 + H_2CO_3 = Ca^{2+} + 2HCO_3^-$$
(6.6)

or dolomite dissolution:

$$MgCa(CO_3)_2 + 2H_2CO_3 = Mg^{2+} + Ca^{2+} + 4HCO_3^{-}$$
(6.7)

These reactions contribute Mg^{2+} , Ca^{2+} and HCO_3^- to solution and they are accordingly important components of the chemistry at location **b**. The groundwater at this location is relatively fresh, with an EC about 2,000 μ S/cm, and so small contributions of ions from water-rock interaction can easily change the chemistry.

The same can be said for the groundwater at location \mathbf{c} , which has an EC value of approximately 1,000 μ S/cm. This water has become less saline than that up-flow due to the mixing with water efficiently recharged through the exposed aquifer. The low concentration of Na⁺Cl⁻ type cyclic salts in the recharge water (see Warrnambool rainfall composition in Figure 6.10) is readily dominated by the products of reaction 6.6 and results



Figure 6.10. Representative groundwater electrical conductivity contours (mS/cm) and chemical composition diagrams in cross section for the Upper Tertiary Carbonate Aquifer $(\mathbf{a}, \mathbf{b}, \mathbf{d}, \mathbf{\hat{k}}, \mathbf{d})$ and the Lower Tertiary Sandy Aquifer $(\mathbf{i}, \mathbf{j}, \mathbf{k}, \mathbf{\&}, \mathbf{l})$ in the Port Campbell Sub-basin. Rainfall chemical compositions at Derrinallum and Warrnambool are also depicted (Hutton & Leslie, 1958). Each chemical composition chart shows the relative abundance of major anions on the left and major cations on the right; labels are indicated on the rainfall charts. in the development of a $Ca^{2+}HCO_{3}^{-}$ type water.

Interestingly, the groundwater composition at the coastal location **d** reverts to Na⁺Cl⁻ type. This water is notably more brackish (EC $\approx 3,000 \ \mu$ S/cm) than that at location **c**. The chemical composition is very similar to that of the coastal rainfall, which in turn is closely related to ocean water composition. The groundwater salinity and type is due to mixing with shallow groundwater, which is more concentrated (EC < 5,000 μ S/cm) possibly because of diffusion of wind-blown salt from the ocean.

Tyrendarra Sub-basin

Figure 6.11 is a cross section showing the schematic changes to the groundwater chemistry along the regional flow path for the Tyrendarra Sub-basin. The Upper Tertiary Carbonate Aquifer has a complex EC pattern in the up-flow area of the cross section which relates directly to the lack of development of regional flow systems in that section of the aquifer. In the southern half of the cross section the coastward regional flow is better developed. In this region, the resident groundwater is gradually less saline along the flow line, which is in direct conflict with the concept of normal evolution of groundwater consisting of the accumulation of solutes. The mechanism for this process is mixing along flow of the relatively brackish water from the centre of the cross section (EC = 4,000 μ S/cm) with water of a fresher composition that is continuously recharged to the exposed aquifer. This mechanism is more influential on the salinity of the water than the opposing mechanism of solutes being gained from water-rock interaction.

The groundwater composition in the Upper Tertiary Carbonate Aquifer at location **a** in Figure 6.11 is similar to inland precipitation chemistry (Figure 6.10) with the difference of additional of chloride, calcium and magnesium. Water-rock interactions described hitherto and some dissolution of halite are the reasons for these changes. Groundwater of a similar chemistry is located at **b**, although the addition of sulphate and loss of magnesium is notable. Relatively less Mg^{2+} is present in this water as it has had less contact with the mafic rocks of the volcanic aquifer. The sulphate is either biological or volcanic in origin (Section 6.3.2).

From location **b** through **c** to **d** the proportion of bicarbonate dissolved in the water gradually increases, although the salinity of the water decreases. The proportion of sodium also decreases at the expense of the combined concentration of magnesium and calcium. All these trends are consistent with the influence of carbonate dissolution (Equations 6.6 and 6.7). The result is that the final groundwater composition at location **d** has a composition markedly different from that of coastal rainfall (Figure 6.10).

This may seem contradictory, given that the groundwater at **d** is shown to be fresh $(\text{EC} \approx 1,500 \ \mu\text{S/cm})$ because of dilution from recharge inputs. However, the chemistry of




the groundwater at **d** differs markedly from rainwater because the relatively small amount of solutes added by water-rock interaction easily have an influence on the composition of water with an already low solute concentration. The groundwater analysed in this study was found to have calcite saturation indices all greater than zero, indicating that further dissolution is not possible. However, recharge of groundwater along the flowpath perpetuates dissolution reactions in the subsurface because of continued introduction of carbonic acid or biogenic acids from the soil zone. As is discussed further in Chapter 7, another likely source of CO_2 is volcanogenic emanations.

Gambier Sub-basin

The groundwater geochemistry within the Upper Tertiary Carbonate Aquifer in the Gambier Sub-basin is presented in Figure 6.12. As with the previous cross sections, the EC of the groundwater is lower in the southern end of the section. The aquifer is blanketed by the Plio-Pleistocene Heterogeneous Aquifer through most of the region and interaction in the form of local flow systems is important in controlling the groundwater salinity in the up-flow areas.

In both the up-flow locations, **d** and **c** (Figure 6.12), the groundwater has a chemical composition that reflects the source precipitation closely (c.f. inland rainfall from Figure 6.10). Both sites do present a slightly higher proportion of calcium in the groundwater than rain water. Addition of this ion can be achieved by weathering of the overlying carbonates of the Plio-Pleistocene Heterogeneous Aquifer, i.e. via reactions 6.6 and 6.7. However, a corresponding increase in the proportion of HCO_3^- , the other product of these reactions, is not present at sites **d** and **c**. Many of the Na/Cl ratio values for water in this area are about 0.7 (c.f. seawater ratio = 0.86; Bear *et al.*, 1999; Figure 6.5), indicating some loss of sodium as chloride is relatively conservative. Therefore, cation exchange between sodium and calcium is the most likely explanation of this chemistry pattern. Although the calcium usually has a higher affinity for sorption, the higher salinity in this water (EC = 2,500 μ S/cm at **d**; EC = 4000 μ S/cm at **c**) creates a greater probability for sodium transition out of solution.

The groundwater composition and chemical evolution at location **b** is considerably different. The groundwater has developed into a $Ca^{2+}HCO_3^{-}$ type water from the Na⁺Cl⁻ type of the source rainwater and is less saline than the groundwater at **d** and **c**. This is the result of more rapid infiltration of recharge water, avoiding evaporation, and subsequent dissolution of cabonates from the host aquifer and the overlying younger units. It is likely that both calcite (Equation 6.6) and dolomite (Equation 6.7) dissolution are progressing within the limits of saturation, although magnesium has the lesser proportion in the groundwater. Magnesium may be lost from solution by cation exchange.





The groundwater composition at **b** persists unchanged until the coastal zone, where mixing with ocean water drastically changes the water chemistry. This mixture is represented by location **a** where the groundwater has an EC value in excess of 30,000 μ S/cm (c.f. seawater EC $\approx 50,000 \,\mu$ S/cm). The chemical composition at **a** is once again Na⁺Cl⁻ type, almost identical to coastal rainfall (Figure 6.10), which in turn is derived from the proportion of solutes in the ocean. This corroborates with other evidence to confirm the cause of the salinity to be seawater intrusion (Chapter 4).

Further insight into water-rock interaction

To this point, carbonate water-rock interaction has been the main type of interaction discussed for the groundwater of the Upper Tertiary Carbonate Aquifer and only brief reference has been made to inheritance of silicate weathering products from the overlying Pliocene-Quaternary Volcanic Aquifer. An important finding, however, is that evidence of silicate water-rock interaction has been observed, even in locations where overlying volcanic rocks are not present. The evidence comes in the form of the silicon/chlorine ratio which is sensitive to the presence of dissolved silica or silicic acid in the groundwater. Figure 6.13 shows this parameter against the changes in the sodium/chlorine ratio which is also a good indicator of water-rock interaction. The values of the groundwater samples can be compared to the ratios expected for rainwater (coastal rainfall has a similar Na⁺/Cl⁻ ratio to the ocean).

In the groundwater from the Upper Tertiary Carbonate Aquifer both ratios are often higher than the rainfall compositions. With the assumption that chloride acts conservatively, this observation reveals the addition of sodium and silicon to solution. Interestingly, some of the Upper Tertiary Carbonate Aquifer samples have higher silicon/chlorine ratios and lower Na^+/Cl^- ratios than some of the samples from the Lower Tertiary Sandy Aquifer. These cases are mostly samples from bores in the region of Mt Gambier and the ash deposits of the region are the cause of this geochemical pattern. The ash is composed partly of pyroclastic glass of very small grain size. The reactive nature of glass (amorphous silica; SiO₂) and the large surface area due to the grain size act to enhance the dissolution of silicon in the groundwater.

Other elements may also give insight into water-rock interaction. All but a few samples taken from this aquifer in this study have no detectable iron concentration. Some dissolution of siderite (FeCO₃) may be present, but subsequent iron oxidation, for example to form goethite, may be the reason for low iron concentrations. No lithium or aluminium were detected in the groundwater samples. This is to be expected for aluminium as it is highly insoluble at the surface and will remain in the solid phase by forming clay minerals (e.g. Equation 6.4).



Figure 6.13. Silicon-chlorine and sodium-chlorine molar ratios of groundwater sampled in this study. A rainfall silicon-chlorine ratio range (0.02-0.04; Gibbs, 1970; Probert, 1976) and the oceanic sodium-chlorine ratio of 0.86 (Bear *et al*, 1999) are labelled for comparison.

Despite the presence of silicate weathering, carbonate water-rock interaction is more important for controlling the chemistry of the groundwater for this aquifer. As was seen in Figure 6.9, the water can be dominated by any of the three major cations. More mature waters in carbonate aquifers tend to accumulate magnesium in the dissolved phase at the expense of calcium, despite which minerals are dissolved, because calcium is preferentially lost through pure calcite precipitation (Leaney & Herczeg, 1995; Herczeg *et al.*, 1997). In other words, magnesian carbonate minerals in the aquifer can be replaced by calcite (Bricker & Garrels, 1967). On the other hand, water of variable salinity in the aquifer is dominated by sodium. This can be the result of both Mg^{2+} and Ca^{2+} exchanging for sodium in the solid phase (clay minerals).

6.3.4 Upper Tertiary Aquitard

The salinity of the groundwater of the Victorian section of the Upper Tertiary Aquitard was mapped using the proxy of EC and is presented in Figure 6.14. There was no data for the South Australian section available for interpolation. On first inspection, Figure 6.14 seems to show that the aquitard hosts more groundwater of a brackish nature (EC > $3,000 \ \mu\text{S/cm}$) than the overlying units (Figures 6.6 & 6.8) and the underlying units (Figure 6.16, 6.18). However, this may be a deceptive aspect of the preliminary map provided in Figure 6.14 because large areas depicted with such conductivity values are in fact devoid of appropriate bores screened in the aquitard. Therefore, doubt still remains as to the groundwater quality in those areas.

In contrast, some areas are well documented and show legitimate salinity patterns with significant variability. In these areas, the distribution of low- and high-EC groundwater in the aquitard closely matches the patterns in EC of the shallow groundwater (Figure 6.6). The reason for the development of these patterns is described in Section 6.3.1. The reason for their development in the aquitard is that the flow systems of the upper levels are linked to those in the aquitard. This connection requires the absence of the Upper Tertiary Carbonate Aquifer, or at least, only a thin presence. Indeed, the aquifer is absent in much of the relevant areas, namely, the north of the Corangamite region and east and north of the Port Campbell Sub-basin. The important implication of this finding is that the factors controlling groundwater quality at the surface are equally important for the quality of water at depth.

The major ion composition of the groundwater in the aquitard is represented in Figure 6.15 by ternary diagrams. The low-salinity water is $Na^+HCO_3^-$ or $Ca^{2+}HCO_3^-$ type and the high salinity water is Na^+Cl^- type. Many of the compositions compare closely to rainwater composition with the notable additions of bicarbonate, calcium and magnesium. These ions are derived from water-rock interaction, namely the reactions of Equations 6.6 and 6.7. Given the close connection with the Pliocene-Quaternary Volcanic Aquifer in regions where the Upper Tertiary Carbonate Aquifer is absent, the weathering of mafic volcanic rocks is also a likely factor (Section 6.3.2).

The geochemistry of the groundwater of the Upper Tertiary Aquitard is not discussed in terms of along flow evolution in this study because of inconsistent spatial distribution of the existing records. The Victorian records' locations are marked by crosses in Figure 6.14; they are numerous in some regions, especially the Corangamite region, but are very rare in others, e.g. southern areas of the Tyrendarra and Port Campbell Sub-basins.

6.3.5 Lower Tertiary Sandy Aquifer

The EC patterns in the groundwater of the Lower Tertiary Sandy Aquifer are obligatorily linked to the recharge patterns and flow systems of the aquifer described in Chapter 4 (Figure 6.16). For example, the recharge zone in the major area of the aquifer's outcrop, around the Otway Ranges, contributes very fresh water to both the Port Campbell and







Figure 6.15. Ternary diagrams showing the relative abundance of the major ions for all Upper Tertiary Aquitard records in Victoria. The relative size of the points relates to the total concentration of ions (meq/L). Precipitation composition (including dry deposition) is represented by two samples from Hutton and Leslie (1958; size of symbols not to scale).

Barwon Downs Sub-basins because of the good hydraulic conductivity of the aquifer and because of the higher P/E ratio of the region due to an orographic precipitation effect.

However, the western part of the Port Campbell Sub-basin contains more saline groundwater (EC up to 9,000 μ S/cm). This zone of elevated EC is contiguous into the region of the aquifer that sits over the Warrnambool High; although only a small number of bores are present to confirm this (Figure 6.16). This area of poorer quality water has previously been mapped (Blake, 1980). Connate water remaining in zones where flux is weak or sluggish has been suggested as the cause of similar variability in quality in this area and elsewhere in the Lower Tertiary Sandy Aquifer (O'Driscoll, 1960; Johns, 1968; Thompson, 1972). Similarly, connate water trapped in the overlying or interfingering low conductivity layers and released slowly by diffusion has also been suggested as a cause (O'Driscoll, 1960). However, subsequent studies rejected the idea (Johns, 1971; Blake, 1980). Further, any contribution of marine connate water as a source of the resident groundwater in the aquifer is refuted by the isotopic evidence presented in Chapter 7.

This issue of variable quality requires further investigation, as the true heterogeneity could not be fully represented in this study either by the map (Figure 6.16) or the data used (water samples from existing wells). Some of the lateral heterogeneity is represented by the map, but in some places the variability is distributed vertically, as observed in deep bores at Portland (pers. comm. Wannon Water, 2005).

It is asserted here that the presence of high-EC water over the Warrnambool High and in the north of the Tyrendarra Sub-basin is a result of infiltration of water of similar poor quality from the aquifer's recharge zones through overlying layers, and ultimately, from the surface (for detail of recharge zones refer to Section 4.8). Upon review of the relative up-flow zones in Figures 6.6 and 6.8, it can be seen that water of comparable EC exists in the overlying aquifers. Naturally, further water-rock interaction along the flow path up to, and within, the Lower Tertiary Sandy Aquifer adds solutes to the water and therefore results in some small increase in the EC values. Additionally, mixing along flow and during vertical recharge acts to homogenise the salinity concentration.

Johns (1968) had already recognised the important role of groundwater-surface water interaction when he noted that high chlorinity at some points in the northern regions of the Lower Tertiary Sandy Aquifer may be partly due to recharge from salty river water and/or areas of poor surface drainage. The interesting aspect of this theory is revealed with the realisation that not all of the water in the Tyrendarra Sub-basin is saline. The water in the lower regions of the flow path is fresh (EC $\approx 1,500 \ \mu$ S/cm), indicating that the recharge to the aquifer sometime in the past was fresher. In the Tyrendarra Subbasin, the salinity near the surface has been developing for at least 8,000 years (Bennetts *et al.*, 2006), whereas periods prior to this, e.g. before the last glacial maximum, are





known have relatively high P/E ratios and low lake salinity (Chapter 3). Therefore, the change in groundwater EC along the flow path in the Tyrendarra Sub-basin illustrates the palaeoclimate fluctuations of the recharge zone.

In contrast, the groundwater EC in the Gambier Sub-basin rarely exceeds 2,000 μ S/cm. The patterns in EC in the Lower Tertiary Sandy Aquifer (Figure 6.16) directly reflect those of the overlying units (Figures 6.6 and 6.8). The controls on groundwater quality in those aquifers are described in Sections 6.3.1 and 6.3.3 and are in turn the controls for the Lower Tertiary Sandy Aquifer's groundwater quality.

The relative concentrations of cations and anions in the groundwater of Lower Tertiary Sandy Aquifer are provided in Figure 6.17. Approximately half of the samples contain chloride as the dominant anion and the remaining half are dominated by carbonates. Fewer than 1% of the samples are sulphate dominated. Both the very fresh and the more saline water of the aquifer are Na⁺Cl⁻ type, whereas the intermediate salinity water tends to be Ca⁺HCO₃⁻ type. Although, even those samples which are dominated by calcium contain a maximum of only 65% Ca²⁺. Magnesium is of lesser importance; all samples have less than 40% Mg²⁺. Enrichment of magnesium tends to be at the expense of calcium rather than sodium.

Despite revealing these patterns, Figure 6.17 does not provide insight into the geochemical evolution of the groundwater along the flow path. Therefore, the cross sections of Figures 6.10, 6.11 and 6.12 are examined in order to gain this insight. One common trend is exhibited in each of the cross sections. All of the sub-basins, except the Warrnambool High region, exhibit a dominance of Cl^- in the upper reaches of the flow path and an increasing importance of HCO_3^- along flow to a point where HCO_3^- eventually becomes the dominant anion. The Cl^- dominant water is associated with sodium and hence is the Na⁺Cl⁻ type mentioned above. The Warrnambool High region hosts mostly Na⁺Cl⁻ type water.

Port Campbell Sub-basin

The hydrogeochemical evolution in the Port Campbell Sub-basin can be divided into two separate trends, the first relating to the flow paths originating in the northeast of the Subbasin and the second for those originating in the northwest. The first evolution pathway is not depicted in the cross section of the Port Campbell Sub-basin (Figure 6.10) and so it is briefly described below, before addressing the cross section in more detail.

As explained in Section 4.8, recharge in the east occurs via direct infiltration in the region of Aquifer exposure and the result is the very low salinity water of that area (Figure 6.16). The water is Na^+Cl^- type in the recharge zone and further along the flow path, for example near Port Campbell, it has evolved into $Na^+HCO_3^-$ type. Throughout this



Figure 6.17. Ternary diagrams showing the relative abundance of the major ions for all Lower Tertiary Sandy Aquifer records in Victoria. The relative size of the points relates to the total concentration of ions (meq/L). Precipitation composition (including dry deposition) is represented by two samples from Hutton and Leslie (1958; size of symbols not to scale).

evolution the salinity changes very little but the changed chemical composition is a result of water-rock interaction, which is explored further below.

In contrast to this flow path, the evolution along the flow from the northwest of the Sub-basin involves some considerable changes in salinity (Figures 6.16 and 6.10). The Na⁺Cl⁻ type groundwater at the beginning of the flow path (location **i**, Figure 6.10) has an EC already elevated to about 4,000 μ S/cm because of the process of significant evapotranspiration at the surface during recharge. The water-rock interactions that accompanied this recharge are detailed in Sections 6.3.2 and 6.3.3 above and would have delivered sodium, magnesium, calcium and bicarbonate to solution, amongst other ions. Subsequent precipitation of calcite is the likely cause of the lower proportions of Ca²⁺ and HCO₃⁻ at location **i** when compared to the proportions in rainfall (Figure 6.10).

Having reached the Lower Tertiary Sandy Aquifer close to location \mathbf{i} , the groundwater undergoes a chemical change within the aquifer that is represented by the composition at locations \mathbf{j} and \mathbf{k} further along the flow path. The proportion of major ions in solution at \mathbf{j} and \mathbf{k} is almost identical (Figure 6.10) and the salinity at the two sites is also similar, although higher at \mathbf{j} than at \mathbf{k} . The changes that have occurred in the water since location \mathbf{i} are a relative increase in sodium at the expense of magnesium and the addition of bicarbonate. The reactions responsible for these changes are sodium-magnesium cation exchange and silicate water-rock interaction (e.g. Equation 6.4; albite or other sodic feldspar comprises up to 5% of the aquifer's matrix; Baker, 1961). Incongruent silicate dissolution is supported by the increase in the silicon/chlorine ratio concurrent with an increase in the sodium/chlorine ratio (Figure 6.13). The increase in sodium proportion as a result of cation exchange has been previously identified (Johns, 1968; Blake, 1980). Some minor interaction with carbonate minerals within the aquifer is also probable.

These same processes are responsible for the development of Na⁺HCO₃⁻ type water further along flow, at location l (Figure 6.10). However, the important difference at this location is that the salinity is much lower (1,000 μ S/cm). This allows the addition of solutes from the reactions to influence the composition more easily. It also indicates, assuming that the fresh water is older than that at locations **j** and **k**, that the quality of the groundwater that has been recharged to the aquifer was better in the past. The accumulation of salinity at the surface in the recharge zone due to climate change has caused an apparent decrease in salinity along the flow path.

Formation of Na⁺HCO₃⁻ type water via water-rock interaction has been identified in other basins, e.g. the Atlantic Coastal Plain (Foster, 1950). For the development of this type of water, with such high concentrations of bicarbonate, a source of dissolved CO₂ other than soil CO₂ was found necessary; this source was deemed to be subsurface production from decomposition of organic matter in the aquifer (Foster, 1950). Concentration of alkalinity above that expected from input of soil CO₂, i.e. 7 meq/L, is observed in some samples from the Lower Tertiary Sandy Aquifer from each of the sub-basins of the Otway Basin (Johns, 1968; this study). Johns (1968) attributed the additional CO₂ to a minor bacterial input (e.g. sulphate-reducing bacteria) and a more important input from *in situ* decay of carbonaceous matter. However, it is more likely that the important contribution of additional CO₂ to the aquifer is from volcanic gas. Such contributions are known to result in the formation of Na⁺HCO₃⁻ type water due to the aggressive nature of the acidic water produced (D'Amore & Panichi, 1985). Volcanic inputs are discussed in more detail in Chapter 7.

Tyrendarra Sub-basin

The evolution of the Lower Tertiary Sandy Aquifer groundwater along flow in the cross section of the Tyrendarra Sub-basin (Figure 6.11) is very similar to that described above for the Port Campbell Sub-basin groundwater. The water originates as Na⁺Cl⁻ type, a composition inherited from a rainfall signature and concentrated by evapotranspiration (location **i**). It also has a considerable proportion of magnesium derived from volcanic rock weathering and/or dolomite dissolution that has occurred as the groundwater infiltrated through the overlying aquifers. The groundwater's chemistry evolves along the flow path with the addition of bicarbonate and cations from water-rock interaction and subsequent cation exchange that allows sodium to dominate (location **j**). These reactions progress to a point were formation of a Na⁺HCO₃⁻ type water occurs (location **k**).

Another pattern that the Tyrendarra Sub-basin has in common with the Port Campbell Sub-basin is the decrease in salinity along the flow path, as illustrated by the EC contours in Figure 6.11. The differing concentration of salinity at differing locations along the two flow paths and the respective groundwater velocity in the sub-basins (Section 4.8.5) indicates that the timing of salinisation of surface water and shallow groundwater was different in the separate regions.

Gambier Sub-basin

The chemical evolution of the groundwater in the Gambier Sub-basin is markedly different from the regions to its east because of its lack of a basalt cover of the Pliocene-Quaternary Volcanic Aquifer. As discussed above, the effect on drainage of the volcanic rocks and their clay weathering products has led to the concentration of salinity in recharging water. The Gambier Sub-basin contains mostly fresh water because of the effective deep drainage facilitated by the carbonaceous cover of the Plio-Pleistocene Heterogeneous Aquifer. Accordingly, the groundwater of the underlying Lower Tertiary Sandy Aquifer has low salinity (Figures 6.16 & 6.12). As a result, the chemical evolution from water-rock interaction is more easily observed in the more dilute water of the up-gradient areas of the Gambier Subbasin than it is in the brackish water of the other sub-basins. This provides an insight into what the composition of the water in the other areas was like prior to salinisation.

It is important to note that the overall evolution from Na^+Cl^- type to $Na^+HCO_3^$ type that is observed in the other sub-basins is present in the Gambier Sub-basin despite the lower salinity. A minor difference is the lack of a large proportion of magnesium at location l in Figure 6.12 (c.f. location i in Figures 6.10 and 6.11). This indicates that the dolomite in the Gambier Sub-basin is not as effective as the olivine, pyroxene and feldspar of the basalt to the west at delivering magnesium to solution.

The main difference, however, lies in the water composition at the intermediate stage (locations **k** and **j**, Figure 6.12). As the groundwater flows through the Lower Tertiary Sandy Aquifer the relative concentrations of bicarbonate and calcium increase to the point where a $Ca^{2+}HCO_{3}^{-}$ type water is produced. The reactions contributing to these changes include the dissolution of carbonate minerals and incongruent dissolution of silicate minerals in the aquifer matrix driven by dissolution of volcanogenic CO_{2} . The groundwater chemistry, particularly at **j**, is also influenced by mixing with groundwater leaked from the Upper Tertiary Aquitard and the Upper Tertiary Carbonate Aquifer. The mixing is evidenced by the inferred groundwater flow lines and the similarity of the groundwater composition in both aquifers (c.f. **c** and **j**). The water in the overlying layers can be more saline than the Lower Tertiary Sandy Aquifer water (Figure 6.12). Therefore, the leakage need not be voluminous in order to produce a compositional change in the fresher groundwater.

Between locations \mathbf{j} and \mathbf{i} (Figure 6.12), the water-rock interaction proceeds and the added reaction of cation exchange increases the proportion of sodium in the water. The result is the Na⁺HCO₃⁻ type groundwater at \mathbf{i} . Some sodium in solution is consumed in this region to form montmorillonite from kaolinite (Equation 6.5), as shown by the activity diagrams of Blake (1980). However, continued cation exchange acts to maintain Na⁺ as the dominant cation in solution (Blake, 1980). The ability of the reaction of Equation 6.5 to proceed to the right hand side is evidenced by the groundwater chemistry in the Lower Tertiary Sandy Aquifer in a region close to location \mathbf{i} . The samples from this region have no detectable dissolved silicon, even though samples up-flow have silicon/chlorine ratios and silicon concentrations that indicate silicate dissolution (Figure 6.13). The groundwater in this area also has very high pH, with many samples over pH 9. The high pH means a low concentration of H⁺, which drives the reaction to the right (Blake, 1980). The consequent removal of H₄SiO₄ is evidenced by the low silicon concentrations and silicon/chlorine ratios.

is another region in the central Port Campbell Sub-basin. The high pH is supported by the reduced environment in the aquifer, which in turn is the result of volcanogenic inputs and reactions catalysed by microorganisms.

6.3.6 Upper Cretaceous Aquifer

The publicly available EC records for the Victorian section of the upper part of the Upper Cretaceous Aquifer were interpolated to create the map in Figure 6.18. In addition, published values from other sources were collated and are presented in Table 6.2. This table also lists the formations for which EC measurements appear in the map. The map depicts data from the upper part of the aquifer only, namely: the Pebble Point Formation, Moomowroong Sand Member, Timboon Sand and Paaratte Formation, due to paucity of data for the lower part and because of the different salinity of the groundwater from the various formations.

The salinity in the upper part of the aquifer is generally low according to the records used in this study (Figure 6.18). However, samples from the older geological formations, the Paaratte Formation and the Nullawarre Greensand, have yielded salinity values ranging from close to rainfall concentration to well over half the concentration of seawater (Table 6.2; Leonard, 2002). Some of this variability occurs vertically, with higher concentrations usually, but not always, at deeper depths.

The variability is also distributed laterally and the Paaratte Formation groundwater has a pattern of increasing salinity with increasing distance from the aquifer's exposure in the Port Campbell Sub-basin (Leonard, 2002). This pattern is visible to some extent in Figure 6.18 where EC rises to over 3,000 μ S/cm over the Warrnambool High. Given the flow regime of this Aquifer established in Chapter 4, most of the water in this aquifer can be considered to have recharged before or around the same time as groundwater of the overlying Lower Tertiary Sandy Aquifer. Therefore, some of the younger, more saline water observed in the overlying units is yet to infiltrate the Upper Cretaceous Aquifer. For the fresh water sampled from the Upper Cretaceous Aquifer in this study, the source of the water is firmly linked to precipitation by O and H isotopes (Chapter 7). The implication is that this fresh water could have been recharged during the intermittent pluvial periods of the stadials and interstadials prior to the last glacial maximum.

The salinity of the lower part of the aquifer, i.e. Flaxman Formation and the Waarre Formation, can be considerably higher than any values of salinity observed in the Lower Tertiary Sandy Aquifer or the Upper Tertiary Aquitard. As with the upper part of the aquifer, salinity usually increases with depth and is lower closer to the recharge zone of the Otway Ranges (Leonard, 2002). Mehin (1999) reported that high salinity water lies juxtaposed with low salinity water in the Waarre Formation. The reason for this

Formation	${f EC}$ $(\mu {f S/cm})^{-a}$	$\frac{\text{TDS}}{(\text{mg/L})^{-b}}$	$ ext{TDS}$ (ppm) c	${f Na^++Cl^-}\ {f (ppm)}^{d}$
Pebble Point Formation*	140-2,100			
Moomowroong & Wiridjil Formations*	95 - 500	_	_	_
Timboon Sand*	569 - 634		4,590	
Paaratte Formation*	140 - 1,295	325 - 15,793		
Nullawarre Greensand	—	626 - 21,259		
Flaxman Formation		2,048-22,273		
Waarre Formation	—	$1,\!285\!-\!26,\!300$	17,209	400-20,000

Table 6.2. Salinity ranges of groundwater from the upper and lower parts of the Upper Cretaceous Aquifer. *a*—this study: Victorian groundwater management system data; *b*—Leonard, 2002; *c*—Johns, 1971; *d*—Mehin, 1999.

*—records from these units are depicted in the map in Figure 6.18

was thought to be displacement of aquifer blocks due to faulting (Mehin,1999). Some of Mehin's values (Table 6.2) are estimates and others were calculated from electromagnetic logs or measurements on samples from drill stem tests. Oil and gas shows from wells in the basin are usually associated with high salinity formation waters and this is in agreement with the idea that meteoric flushing can diminish the likelyhood of oil and gas remaining in traps (Wopfner *et al.*, 1971). Saline formation water is known to have migrated with hydrocarbons from depth via faults and fractures (e.g. Cai *et al.*, 2001). Therefore, such migration could be the source of the saline water in the Waarre Formation.

Another source that has been suggested is connate water (Wopfner *et al.*, 1971). A saline connate source is possible given the marine origin of the Upper Cretaceous strata. If this is indeed the source, the concentration of the connate water has obviously been diluted by meteroic water in parts of the Formation, as evidenced by the the salinity values below seawater concentration (approximately 35,000 ppm; Table 6.2). This situation is significantly different from the formation of more saline groundwater, such as Canadian shield brine or similar brine elsewhere, which is shown to have developed from seawater that was concentrated prior to mixing with meteoric influx (Cai *et al.*, 2001; Bottomley *et al.*, 2003; Greene *et al.*, 2008).

An alternative explanation to a connate source of the salt is the accumulation of dissolved solids in the groundwater via water-rock interaction over long residence times and possibly evaporitic concentration prior to infiltration. The groundwater of the lower part of the Upper Cretaceous Aquifer was not sampled as part of this study, hence, isotopic evidence relating to the sources is not yet available. Therefore, the provenance of the water and its solutes remains to be determined with any certainty.





6.3.7 Lower Cretaceous Basement

The salinity of water from the Lower Cretaceous Basement of the Otway Basin (the Otway Group) is not widely documented. Even where it is documented, many published values are only estimates. One range of total dissolved solids (TDS) for the basement is 10,000–27,000 ppm (Wopfner *et al.*, 1971). A similar range of salinity, gained from air lift samples from two bores near Port Campbell, is 17,320–26,200 mg/L (Leonard, 2002). Another TDS value, from a sample taken from the Otway Group in 1966, is 19,166 ppm, which lies within the ranges cited above (c.f. 35,000 ppm for typical ocean water; Johns, 1971).

These high salinity values are thought to possibly be connate water (Wopfner *et al.*, 1971). However, it is not possible that the water of the Otway Group is true connate water due to the non-marine (fluvial) origin of the Group's strata (Chapter 2). One possibility is infiltration of marine water during the Upper Cretaceous due to transgressions, or infiltration of Cretaceous marine water from the Upper Cretaceous rocks at a later time.

In contrast, the groundwater of the Basement where it is uplifted in the Otway Ranges is usually fresher. A minimum concentration of Na^+Cl^- of 1,000 mg/L has been reported for one bore and in several cases the salinity increases markedly with depth (Leonard, 2002). This pattern seems to represent an older resident water high in salinity mixing with younger downward-infiltrating meteoric water. As with the Waarre Formation, understanding of the ultimate provenance of the water and its solutes requires further investigation.

6.4 Groundwater quality for beneficial use

6.4.1 Nitrogen

Nitrogen is an important nutrient in groundwater, but in excessive concentration it can cause eutrophic conditions detrimental to groundwater-dependent ecosystems. In drinking water, 50 mg/L of nitrate (NO_3^-) has adverse side effects, and as nitrite (NO_2^-) as little as 0.2 mg/L on a long term basis can be troublesome (World Health Organisation guidelines).

From the 1,950 bores in the Victorian records that exhibit some nitrogen concentration, only 12 bores had concentrations over 50 mg/L (n.b. it is not explicitly stated, but it is assumed that nitrogen concentrations from the records are expressed as nitrate; data are available in Appendix G). The higher concentrations of nitrate were recorded in samples from shallow bores screened in the surficial aquifers: the Plio-Pleistocene Heterogeneous, Pliocene-Quaternary Volcanic and the Upper Tertiary Carbonate Aquifers (Figure 6.19).

This relationship indicates the source of excess nitrate is from the surface. In general, sources of nitrogen dissolved in groundwater are: soil zone nitrate from microbially-



Figure 6.19. NO_3^- concentration in groundwater from Victorian state records vs. depth of bottom of screen interval.

mediated reactions that fix atmospheric nitrogen (N_2) and convert it to nitrate by nitrification; human waste (e.g. septic tank leaks); manure from stock; and fertiliser.

Studies of the nitrate concentration in the Pliocene-Quaternary Volcanic Aquifer revealed that high nitrate could be found in relatively low-salinity groundwater (Lawrence 1983; Bayne, 1996). The presence of such water was found to be generally coincident with land cleared of native vegetation and used for cattle and sheep grazing. The source of nitrate is considered to be from both stock manure and enhanced soil nitrate due to nitrogen-fixing bacteria living in a symbiotic relationship with the clover pasture that replaced the native vegetation (Lawrence 1983; Bayne, 1996). It is also likely that some localised contamination from fertiliser or septic systems has occurred. Nitrogen isotopes could be useful in future studies as an environmental tracer to delineate between biological and anthropogenic sources of nitrate in the groundwater, especially in the aquifers not previously addressed.

6.4.2 Fluorine, Boron and Phosphorous

Fluoride concentrations in drinking water exceeding 1.5 mg/L (0.08 mmol/L) are identified as hazardous to human health by the World Health Organisation. Figure 6.20 shows that two samples taken during this study from the Lower Tertiary Sandy Aquifer, a significant drinking water resource for the region, exceed the flouride limit and most of the waters have significant concentrations of flouride, given their overall low salinity.

The source of fluoride in groundwater is often attributed to the dissolution of fluorinebearing minerals such as fluorite (CaF₂) and fluorapatite (Ca₅(PO₄)₃F) (e.g. Ampabire *et al.*, 1997). These two minerals may be found in sedimentary basins, either as detrital grains from metamorphic or igneous complexes or as diagenetic or hydrothermal cements.

If congruent dissolution of fluorapatite were the main source of F^- to solution, three moles of phosphate would be added with every mole of fluorine. Only 7 of the 87 samples analysed in this study contain any phosphate, and these are not the samples with the highest fluoride content. As there is no positive relationship between the ions, they cannot be attributed to the same source, unless phosphate removal has occurred subsequently. Aside from the unlikely mechanism of precipitation, phosphate removal mechanisms are lacking. Therefore, the dissolution of fluorapatite can be ruled out as a source for these ions. The more plausible source of the arbitrary distribution of phosphate is agricultural contamination.

Similarly, there is no strong positive relationship between molar concentrations of fluoride and calcium in the groundwater to indicated fluorite dissolution (see Figure 6.20). Fluorite dissolution delivers calcium and fluoride to solution at a molar ratio of 1:2. This correlation is not visible in the data of Figure 6.20 for two reasons. The first is that many other, more abundant sources of calcium are contributing to the concentration in groundwater. The second is that, unlike phosphate, Ca is easily lost from solution in fresh waters such as these via sorption and ion exchange. Therefore, conclusive evidence to accept or disregard fluorite as the source of F^- in the groundwater is yet to come to light.

In India, the main primary source of fluorine in groundwater is weathering of F-rich minerals other than fluorite and fluorapatite, i.e. micas, hornblende and pyroxene (Jacks *et al.*, 2005). Once liberated from the mineral matrix, F^- is either directly contributed to solution in groundwater or can be recycled via mobilisation as dust or as a dissolved constituent of rain, forming an important secondary source. When dissolved in groundwater, F^- solubility is governed by fluorite (CaF₂) saturation and by connection, the Ca²⁺ concentration of the water (Jacks *et al.*, 2005; Chae *et al.* 2007). Some of these minerals are present in the Otway Basin, mainly in the basalt of the Pliocene-Quaternary Volcanic



Figure 6.20. Molar concentrations of fluoride plotted against those of boron and calcium in groundwater samples from this study. Dashed lines indicate the maximum acceptable concentrations of F^- and B^{3+} in drinking water.

Aquifer; this source of F^- is linked to the volcanic or magmatic source below.

An important source of F^- in groundwater is volcanic degassing, either via direct dissolution of HF_{gas} , dissolution of F remaining in volcanic or plutonic rocks, or by incorporation of atmospheric precipitation (wet or dry) after degassing (D'Alessandro, 2006; Bellomo *et al.*, 2007). In Korea, F^- concentration in groundwater from plutonic host rocks (granitoids) is significantly greater than that from volcanic rock aquifers (Chae *et al.* 2007). Therefore, although the some basaltic rocks of of Victoria are known to contain fluorapatite (Matsumoto *et al.*, 2000), it may be assumed to be an insignificant source given the low concentrations in water from that aquifer (Figure 6.20). Direct dissolution of HF_{gas} is a more likely pathway for magmatic fluorine to enter the groundwater in the Otway Basin.

Magmatic degassing contributing F^- to deep geothermal groundwater systems has also been associated with addition of boron (Guo *et al.*, 2008). Figure 6.20 shows a positive relationship between these two solutes in a subset of the groundwater samples of this study, most of them being from the deep Lower Tertiary Sandy Aquifer. Boron is also a health risk when present in drinking water over the concentration of 0.5 mg/L (approximately 0.05 mmol/L); 10 samples from this study exceed this limit (Figure 6.20). Despite some samples that contain either B^{3+} or F^- exclusive of the other solute, this data is a good indication that volcanic degassing is the source of fluoride and boron in the groundwater.

6.4.3 Sulphur

Although a toxin if inhaled in gaseous form (H₂S), sulphur dissolved in groundwater is not a health risk to humans. Australian and New Zealand drinking water standards cite maximum concentrations of 0.05 mg/L and 500 mg/L for dissolved H₂S and SO_4^{2-} respectively for aesthetic reasons relating to taste and odour.

The sulphur speciation and concentration in the Otway Basin groundwater is interesting for two reasons. The first, also noted by Johns (1968), is that groundwater samples have SO_4^{2-}/Cl^- ratios both greater and lesser than the oceanic ratio (which is similar to the near-coast rainfall ratio) in a distribution that is not linked to other chemical processes identified in the basin. This implies that mechanisms for both addition and loss of sulphate are at work, and that progressive water-rock interaction is not the most important process. Secondly, the sulphur system is subject to the opposing bacterial reactions sulphate reduction and sulphide oxidation, controlled by redox conditions in the aquifer.

The main processes that contribute sulphur to groundwater of the Otway Basin groundwater are: addition of cyclic salt (input as SO_4^{2-}), bacterial oxidation of pyrite (input as SO_4^{2-}), and magmatic degassing of hydrogen sulphide (input as H_2S). Inputs of SO_4^{2-} can be subsequently converted to H_2S or HS^- by bacterial reduction.

The first contribution of sulphur to the groundwater is cyclic salt, incorporated from rain or dry deposition by recharge water. As a benchmark for comparison, the SO_4^{2-}/Cl^{-} ratio of inland rainfall near the aquifers' recharge zones is 0.036 (Hutton & Leslie, 1958). Groundwater samples from this study exhibit ratios both above and below this benchmark, in the range 0.01–1.23, due to input of other sulphur and also due to redox conditions causing speciation changes.

One potential source of additional sulphur is pyrite (FeS₂), as it is present in the Upper Tertiary Carbonate and Lower Tertiary Sandy Aquifers (Chapter 2). Aerobic bacterial action can break down pyrite into oxidised iron and sulphate, at the same time releasing acid into solution. Some neutral or slightly acidic samples of groundwater have significant iron concentrations. Samples of a similar pH have significant sulphate and dissolved oxygen concentration (Figure 6.21). This indicates that oxidising conditions do result in some pyrite oxidation and release of ions to solution, increasing the SO_4^{2-}/Cl^- ratio and causing acidification. Acidity is subsequently buffered by water-rock interaction.

Another source of additional sulphur is magmatic degassing, a product of Recent volcanism in the Otway Basin. This input is in the form of H_2S_{gas} , which can be dissolved in the groundwater at low to neutral pH. If this additional sulphur remains as dissolved H_2S , it does not change the SO_4^{2-}/Cl^- ratio. However, if the redox conditions allow, oxidation to sulphate will occur (Figure 6.22) and cause an increase in the ratio. This is the case



Figure 6.21. Concentration of dissolved oxygen and dissolved sulphate ions with variation in pH.

for aquifers such as the Upper Tertiary Carbonate Aquifer, whose groundwater exhibits high concentration of both dissolved oxygen and sulphate (Figure 6.21). Further evidence regarding magmatic degassing is discussed in Section 7.2.1.

The explanation of $\mathrm{SO}_4^{2-}/\mathrm{Cl}^-$ ratios of some groundwater samples being lower than the rainfall ratio requires removal of sulphate, assuming that chloride is conservative. The speciation of sulphur is controlled by redox reactions and pH conditions (Figure 6.22). Anaerobic conditions in the aquifer are conducive to sulphate reduction by bacteria. SO_4^{2-} is reduced to sulphide by sulphate-reducing bacteria; the sulphide then forms HS^- (if pH > 7) or H₂S (if pH < 7) in solution (Figure 6.22). Sulphate reduction can therefore lower the $\mathrm{SO}_4^{2-}/\mathrm{Cl}^-$ ratio below original values inherited from rainfall (Johns, 1968). Reducing conditions are present in much of the deep basin, e.g. the Lower Tertiary Sandy Aquifer, and are coincident with high pH and low concentration of both SO_4^{2-} and DO (Figure 6.21). The reduced conditions are perpetuated by interaction with volcanic inputs and act to buffer the pH to alkaline levels.

6.4.4 Other Contamination

The identification of contamination in the groundwater of the Otway Basin and its sources is not the focus of this research. Rather, the physical and chemical hydrogeology framework that has been constructed in this study is designed to aid future investigations as necessary. However, from the understanding thus gained, it is clear that the quality of the



Figure 6.22. Phase diagram for aqueous sulphur species, after Kehew (2001).

groundwater in all the aquifers is vulnerable to conditions at the surface, encompassing land use and surface water quality. In particular, the subaerially exposed karst systems of the basin have been used for waste disposal in the past, which may still be a source of serious contamination (Waterhouse 1977; Shugg, 1993; Vanderzalm & Dillon, 2006).

Future risks of contamination are amplified or complicated by increasing land use relating to intensive agriculture, industry and development in the region. Such ventures in the Otway Basin encompass pine and eucalypt plantations, dairy industry, hydrocarbon exploration, CO_2 geosequestration and geothermal energy. Conscientious attitudes and legislation regarding land management and waste disposal are the best tools for preventing future groundwater contamination. Thorough investigations of risks to deep groundwater quality associated with subterranean ventures are also necessary.

6.5 Synthesis

The shallow groundwater of the Otway Basin has a wide range of salinity, for which the main mechanism of concentration is evapotranspiration. This means the groundwater quality is strongly dependent on the climate and the drainage characteristics of the region. The shallow groundwater is generally fresh in the Gambier Sub-basin because of effective vertical drainage through the surface cover of the Plio-Pleistocene Heterogeneous Aquifer and the Upper Tertiary Carbonate Aquifer. Whereas the Corangamite and Upper Wannon regions have more saline water because of the clay component of the shallow subsurface owing to weathering of the Pliocene-Quaternary Volcanic Aquifer. Because the Port Campbell and Tyrendarra Sub-basins have volcanic cover in the north and exposure of the carbonate or siliciclastic aquifers in the south, the EC of the shallow groundwater appears to decline along the regional flow path. The Gellibrand River region contains mostly fresh water because it has good drainage and a high P/E ratio due of the orographic rainfall over the Otway Ranges.

The solutes of the groundwater in the Pliocene-Quaternary Volcanic Aquifer are sourced from cyclic salt and water-rock interaction. Generally, the high-salinity water is $Na^+Cl^$ or $Na^+Mg^{2+}Cl^-$ type, which is reflective of concentration of cyclic salt with some additional solutes from weathering reactions, and the fresh water is $HCO_3^-Ca^{2+}$ type, which represents the inputs from the fastest reactions within a suite of water-rock interaction that occurs in the aquifer. The reactions include: the dissolution of amorphous silica (from volcanic glass and ash) and olivine; incongruent dissolution of pyroxene and feldspar to form clay minerals such as kaolinite, illite, and smectite, sometimes through an intermediary phase like chlorite; oxidation of iron to form goethite; alteration of clays, e.g. kaolinite altered to montmorillonite; and cation exchange. The first result of these mineral reactions for the chemistry of the groundwater is the addition of metals (i.e. Na^+ , Mg^{2+} , Ca^{2+} and others) and the production of HCO_3^- in solution through acid hydrolysis. Subsequently, dominance of one cation can be achieved through cation exchange and solutes can be lost through mineral precipitation (e.g. calcite) or alteration (e.g. kaolinite to montmorillonite).

The geochemistry of the Upper Tertiary Carbonate Aquifer is controlled to a large extent by the salinity distribution of the shallow groundwater and the reactions within the Pliocene-Quaternary Volcanic Aquifer, both described above. The fresh groundwater of this aquifer especially exhibits addition of calcium and bicarbonate from weathering of volcanic and carbonate rocks. The influence of volcanic silicate mineral weathering is confirmed by concurrent increase in sodium/chloride and silicon/chloride ratios. Saturation indices with respect to calcite are predictably close to or greater than zero in the aquifer, however, there is evidence of ongoing carbonate mineral interaction. This is due to the constant input of soil CO_2 and other organic acids that drive hydrolysis from the surface via local scale flow systems, as well as the introduction of volcanic CO_2 to the system from depth. Cation exchange and calcite precipitation can result in the loss of calcium from solution and the formation of Na⁺Cl⁻ type water in brackish conditions.

As with the Upper Tertiary Carbonate Aquifer, the groundwater quality within the Lower Tertiary Sandy Aquifer is dependent on the quality of water infiltrating from overlying units. Therefore, the groundwater of high to moderate salinity in the aquifer is from meteoric water that was subjected to evaporitic concentration at the surface, rather than from connate water or mixing with connate salt. Hence, in areas where the salinity of the water decreases along the flow path, the changes in climate conditions affecting recharge quality through time are illustrated. These conditions change from place to place, with the Warrnambool High zone containing mostly high salinity Na⁺Cl⁻ type water, indicating the longest history of salinisation in its recharge zone. All the major sub-basins exhibit evolution of the groundwater from Na⁺Cl⁻ type in the up-gradient area, to Na⁺HCO₃⁻ type in the down-gradient zone. The important water-rock interaction reactions are those described above for silicate and carbonate mineral weathering, be they in the form of inherited solutes or *in situ* reaction. Continual water rock interaction, aided by input of volcanic CO₂, and cation exchange along the flow path, sometimes involving alteration of clay minerals, are the causes of the mature water being Na⁺HCO₃⁻ type.

The groundwater of the Upper Cretaceous Aquifer is very different in the upper and lower parts. The upper part contains much fresh water, although salinity can increase with depth. This fresh water is a testament to the meteoric source of the water, and to the idea that poorer quality meteoric recharge has yet to enter the aquifer from above. However, the lower part contains groundwater of more variable salinity with changes often associated with placement of faults. This water may be the result of mixing of meteoric and connate sources, and so is certainly different from other brines around the world. Further research is needed to unravel the evolution of this water. The water hosted in the fractures of the Lower Cretaceous Basement is similar to the lower part of the Upper Cretaceous Aquifer and the ultimate source is still debateable. Salinity can be very low in the Otway Ranges area but usually increases with depth and the water is more saline where the basement is overlain by other units.

The groundwater resources of the Otway Basin, where not concentrated by evapotranspiration, are generally high quality and are not affected on a regional scale by harmful solutes or contamination. Some samples contained harmful levels of fluoride and boron, most of which is derived from volcanic gas. Victorian public records show very few groundwater samples have harmful nitrate concentrations, those that are elevated are near the surface and are associated with human activity. Sulphur in the groundwater of the Otway Basin comes from sources other than cyclic salt, i.e. pyrite oxidation and volcanic gas, and its speciation is controlled by bacterially-catalysed redox reactions.

This chapter provides a base-level understanding of the natural hydrogeochemical evolution in the Otway Basin in order to allow for more effective management of potential risks to groundwater quality that may result from human activity or geological and environmental changes. The chemical evolution defined is in agreement with most published work, strongly linked to the local, intermediate and regional scale flow systems outlined in Chapters 4 and 5, and is shown to be sensitive to climatic changes, such as P/E ratio fluctuations (Chapter 3) and Quaternary volcanic activity (Chapter 2).

Chapter 7

Isotope Hydrogeochemistry

7.1 Introduction

In recent decades advances in isotope hydrogeochemistry have enabled better comprehension of processes affecting groundwater and have therefore resulted in more efficient management of groundwater resources. For example, advances include the development of the understanding of the effects of evaporation and precipitation on the oxygen and deuterium composition of meteoric source waters (Craig, 1961; Gonfiantini *et al.*, 1974; Gonfiantini *et al.*, 1976). Carbon-14 dating of groundwater had begun in the 1950s (e.g. Vogel, 1959) and subsequently studies began to utilise an integrated approach with several isotopic analytes, including the correction or interpretation of radiocarbon groundwater ages aided by ¹³C data (Hufen *et al.*, 1974; Winograd & Pearson, 1976). As analytical techniques developed, isotopes of other important solutes were studied and utilised in diverse ways, such as the application of ³⁶Cl to dating very old groundwater (Phillips *et al.*, 1986) and to detecting modern recharge (Bentley *et al.*, 1982).

Despite this demonstrated applicability, Back and Freeze noted that, in 1983, "isotopes have yet to become an integral part of North American hydrogeologic investigations". Today, the utilisation of isotopic techniques to inform water resource administration does occur (Cook, 2009), but is still not as widespread as some would hope.

Recently, the most sophisticated addition to a hydrogeochemist's toolkit has become reactive transport modelling of isotopic tracers, enabling the calculation of "age mass" (Bethke & Johnson, 2008). This provides a more accurate assessment of groundwater age than previous radiometric piston flow models could provide and is especially applicable in heterogeneous aquifers or fractured rock aquifers (Cook, 2009). Unfortunately, modelling is not attempted in this study. Rather, several stable and radiogenic isotope systems are used to characterise the processes affecting the Otway Basin groundwater, building upon knowledge gained from prior investigations.

7.1.1 Previous Work

Several authors have used isotopic methods of investigation in the Otway Basin. In the Gambier Sub-basin calculations of groundwater velocity and recharge rates have been made using carbon-14 and tritium; in addition to characterisation of the source of the groundwater, the nature of the recharge mechanisms and the processes affecting dissolved carbon through stable isotopes (Allison & Hughes, 1975; Love *et al.*, 1992a; Love *et al.*, 1993; Leaney & Herczeg, 1995; Herczeg *et al*, 1997). The number and scale of existing isotopic investigations in the Gambier Sub-basin is greater than the other regions and therefore the focus of this study is mainly on the eastern Sub-basins.

Carbon dating and groundwater source identification has been performed in the Barwon Downs Sub-basin for the Lower Tertiary Sandy Aquifer (Petrides & Cartwright, 2005). A similar multi-tracer approach has been employed for the Plio-Pleistocene Heterogeneous and Pliocene-Quaternary Volcanic Aquifers of the Corangamite region (Cox *et al.*, 2007). Lastly, the isotopic hydrogeochemistry of the surficial aquifers and the Upper Tertiary Carbonate Aquifer in part of the Tyrendarra Sub-basin was addressed by Bennetts (2005).

7.1.2 Methods

This Chapter presents original data which document the isotopic composition of the groundwater of the different hydrostratigraphic units in the Otway Basin and its solutes. The specific isotope parameters reported for the groundwater samples in this study are: δ^{13} C, δ^{2} H, δ^{18} O, 87 Sr/ 86 Sr, radiocarbon concentration relative to modern atmosphere (%mC) and R³⁶Cl/Cl. The groundwater samples were taken at the same time as those used for major and minor ion concentration analysis. Hence, the method of collection, preservation, preparation and analysis is detailed in the methods section of the preceding chapter (Chapter 6) and in Appendix L.

7.1.3 Isotope Notation

A common term used to describe the isotopic composition of matter is δ . Values of δ represent ratios of certain rare isotopes in comparison to their abundant counterparts and are measured relative to international standards. For example, δ^{18} O is the ratio of ¹⁸O to ¹⁶O in the sample as compared to the international standard, Vienna standard mean ocean water (V-SMOW; see equation 7.1). The VSMOW standard is also used to measure δ^{2} H, which quantifies deuterium and hydrogen ratios. The international standard for δ^{13} C, is V-PDB, the Vienna Pee Dee Belemnite, representative of modern to recent marine carbonates.

$$\delta^{18}O_{sample} = 1000 \frac{{}^{18}O/{}^{16}O_{sample} - {}^{18}O/{}^{16}O_{VSMOW}}{{}^{18}O/{}^{16}O_{VSMOW}}$$
(7.1)

Fractionation of isotopes, or how the distribution of isotopes changes between phases or compounds, can be quantified by three terms, which are linked to δ . These are: the fractionation factor, α ; the difference, Δ ; and the separation, ε . For all systems there is a similarity between these fractionation terms (Equation 7.2). Fractionation can be measured experimentally by determining isotope ratios of the two phases involved after allowing them to reach equilibrium with each other at a given temperature. The ratios are then used to calculate the α , Δ or ε .

$$1000ln(\alpha) \approx \Delta \approx \varepsilon \tag{7.2}$$

The example of ¹⁸O fractionation in H₂O between the liquid water and vapour phases illustrates this well. The α and the Δ for ¹⁸O between water and vapour are shown in equations 7.3 and 7.4 respectively. From these equations it is clear that if Δ is greater than zero, or α is greater than one, the heavy isotope is enriched in the first phase or molecule of the equation. In the case of ¹⁸O in water vs. vapour at 25°C, Δ is 9.3‰. Hence, ¹⁸O prefers the molecules of the liquid phase over the gaseous phase and will exchange with ¹⁶O accordingly, producing the fractionation.

$$\alpha_{[water-vapour]} = \frac{{}^{18}O/{}^{16}O_{water}}{{}^{18}O/{}^{16}O_{vapour}} = \frac{1000 + \delta^{18}O_{water}}{1000 + \delta^{18}O_{vapour}}$$
(7.3)

$$\Delta_{[water-vapour]} = \delta^{18} O_{water} - \delta^{18} O_{vapour} \tag{7.4}$$

The degree of fractionation between certain phases or molecules is affected by the temperature at which the process or reaction takes place. Higher temperatures produce less fractionation. For example, evaporation of water at high temperatures will produce a vapour of similar isotopic composition to that of the water. The same process at cooler temperatures will produce vapour that is significantly different, i.e. isotopically lighter than the water, due to the greater fractionation. This relationship is expressed as the variation of the fractionation factor, α , with respect to the temperature term, T [K] in equation 7.5 (where a, b and c are experimentally derived constants).

$$1000ln(\alpha) = a\frac{10^6}{T^2} + b\frac{10^3}{T} + c \tag{7.5}$$

7.2 Stable Isotope Hydrogeochemistry

7.2.1 Oxygen-18 and Deuterium

Origins of the groundwater

The distribution of heavy oxygen and deuterium in the water molecules of the groundwater of the Otway Basin is depicted in Figure 7.1. The δ^{18} O and δ^{2} H values for the majority of samples range from -6 to -4 and -31 to -18 % VSMOW respectively. A few outliers are more enriched in the heavy isotopes, but generally remain depleted in comparison to the standard mean ocean water (SMOW).

The global meteoric water line is plotted in Figure 7.1 according to the equation of Craig (1961), defined as:

$$\delta^2 H = 8\delta^{18} O + 10 \tag{7.6}$$

In addition, the stable isotopic composition of rain water for Melbourne and Adelaide were obtained from the GNIP database (http://www-naweb.iaea.org/napc/ih/GNIP/IHS _GNIP.html). The data were provided as weighted monthly mean δ^{18} O and δ^{2} H values with corresponding precipitation amounts (mm) for the years 1961 to 2001. The monthly mean δ^{18} O and δ^{2} H values for Melbourne were plotted and a line approximating their distribution was used as the local meteroic water line (LMWL) for the Otway Basin (Figure 7.1). Annual weighted means were calculated for each of the cities and are plotted as points to indicate the expected average composition of recharge water (Figure 7.1).

Variation of O and H isotopes in precipitation is controlled by fractionation during cloud formation and condensation of rain droplets. Several factors regarding these processes affect the degree of fractionation, including relative humidity, temperature and amount of precipitation. In general, a hot climate, first or small rain events and low latitudes produce enriched precipitation. The different isotopic compositions relating to variable climate are useful for examining not only whether the source of groundwater is meteoric, but also if it is relatively modern. The Melbourne weighted annual average precipitation has a more depleted composition than that for Adelaide. This is due to the more arid climate in Adelaide and the lower latitude leading to hotter temperatures there, producing less fractionation. Values of δ^{18} O and δ^{2} H in precipitation in the Otway Basin do not exist, however, rainfall composition of the region is likely to be similar to that of Melbourne, with a slight depletion with increased distance from the coast. Further, precipitation in the west of the basin, which lies at lower latitudes than the eastern part, may be more enriched in heavy isotopes due to the latitude effect.

The majority of groundwater samples from this study lie close to the GMWL and



Figure 7.1. Oxygen (δ^{18} O) and hydrogen (δ^{2} H) isotope composition of all groundwater samples. Two weighted annual average precipitation values are provided: the more enriched is for Adelaide, the more depleted for Melbourne. The local meteroic water line (LMWL) is derived from monthly weighted Melbourne data. The processes in the inset, top left, were modified from D'Amore & Panichi (1985)

LMWL (Figure 7.1) and hence can be linked to a meteoric origin rather than connate or marine water. However, some there are some outliers from the main dataset. The outlier of the Upper Cretaceous Aquifer is dismissed as unreliable because the correction applied to it (Appendix L) was formulated from more depleted samples.

The two outliers from the Plio-Pleistocene Heterogeneous Aquifer lie displaced to the right of the LMWL and are enriched compared to the weighted average precipitation points (Figure 7.1). A line connecting either of these points to the annual average Melbourne precipitation has a slope of 5.3 or 5.6. A trend with a slope of 4 to 6 on a δ^{18} O and δ^{2} H plot strongly indicates enrichment of heavy isotopes due to evaporative fractionation (Kehew, 2001). Further evidence that the groundwater from these two sites in the Plio-Pleistocene Heterogeneous Aquifer has been affected by evaporation is the elevated electrical conductivity, approximately 10,000 μ S/cm. One of these samples was taken from a well close to the salty upper reaches of the Hopkins River near Lake Bolac (EC of 17,000 μ S/cm) and

the other was taken from a well situated about 10 km from Lake Corangamite (EC greater than 200,000 μ S/cm). This is an example of surficial, local groundwater flow paths both influencing and being influenced by surface water systems. Hence, the process of evaporation, which is important in the unsaturated zone, is proven to be significant for the shallow groundwater of the Otway Basin. This is confirmed by more numerous samples from saline environments in the Corangamite region by another study (Cox *et al.*, 2007).

The final outlier in Figure 7.1 is a groundwater sample from the Upper Tertiary Carbonate Aquifer taken near Port Macdonnell (δ^{18} O: -1.48; δ^{2} H: 1.3). This water was strongly suspected to have been affected by sea water intrusion (Chapter 4). Being so comparable to the SMOW value (0, 0), these values are confirming evidence that the rise in salinity in the water of this bore (CAR 061) is due to marine incursion.

Aside from outliers, the groundwater samples are tightly clumped and lie in a trend similar to the slope of the meteroic water lines (Figure 7.1). Therefore, it is deduced that the variation along this trend represents the natural variability of the meteoric source water. First, as predicted, there is a weak negative relationship between longitude and δ^{18} O, meaning that samples in the west have a greater proportion of heavy O and H isotopes. This is a result of the slightly drier and hotter climate in the west of the basin compared to its east. Second, many of the samples from the east are more depleted than the average weighted annual precipitation for Melbourne. Some workers might attribute this to colder climates in the past and dub the samples palaeowater, as is the case for groundwater in other parts of the world (Darling *et al.*, 1997; Zongyu *et al.*, 2003). However, the most depleted water in the Lower Tertiary Sandy Aquifer was sampled close to the regional recharge zones and is therefore not the oldest water in the aquifer. Hence, the strongly depleted signature is more likely a testament to the current isotopic composition of precipitation and to the fact that most water reaching the aquifer is sourced from winter rains; as evaporation in the summer months prevents significant deep drainage.

Despite what the known variability of the climate of the Otway Basin region (Chapter 3), spatial analysis of the δ^2 H and δ^{18} O results did not show any trend along groundwater flow paths. It is therefore concluded that palaeoclimate shifts are not visible in this dataset.

Thus, the origin of the groundwater has been confirmed as meteoric, with most samples closely resembling modern rainwater. However, the main group of groundwater samples is offset to the left of the LMWL (Figure 7.1). This implies that during percolation of rainfall through the unsaturated zone, or since recharge to the aquifers, a process has affected the groundwater to fractionate the oxygen and/or hydrogen isotopes. Either depletion of ¹⁸O, enrichment of deuterium, or both has occurred. The processes that are known to cause such fractionation in groundwater are silicate hydration, CO_2 exsolution, H-bearing gas exsolution, or other water-rock interation at low temperatures (top-left inset of Figure 7.1;

DAmore & Panichi, 1985).

Silicate mineral interaction

Silicate hydration occurs in the Otway Basin in the form of production of clay minerals by incongruent dissolution of feldspar and other silicates, as discussed in Chapter 6. Compton *et al.* (1999) showed that an increased degree of hydration of silicates, i.e. weathering of volcanic glass and formation of clay minerals such as smectite and bentonite, resulted in an enrichment of ¹⁸O in the new clay minerals. The corresponding depletion of heavy oxygen atoms occurs in the diagenetic formation fluid, in this case, the groundwater. Additionally, observations of the fractionation of hydrogen in similar systems has revealed that ²H preferentially remains in water molecules, rather than forming the hydrated silicate mineral (D'Amore and Panichi; 1985). This amounts to an enrichment of deuterium in the water due to the loss of hydrogen to the hydration process.

Silicate mineral weathering by the groundwater has therefore fractionated the oxygen and hydrogen isotopes in such a way that displacement of the groundwater data points to the left of the LMWL is expected. However, it is unlikely that this process is the sole cause of the entire fractionation observed (i.e. the displacement in Figure 7.1) because the number of water molecules in a given sample that were probably affected by the water-rock interaction is small compared to the total number of water molecules in a sample. Further evidence that silicate weathering is present but not dominant is provided by the strontium isotope data (Section 7.2.3).

Interaction with magmatic CO_2 and H_2S

Given the geological setting, loss of CO_2 and H_2S is also a viable candidate for causing fractionation of the groundwater. Evidence of the presence of these gases was often detected during sampling, e.g. the distinctive "rotten" scent of H_2S . Some samples, after being bottled without an air bubble, presented gas volatilisation within a matter of minutes. Gas seeps are difficult to detect from dry land without monitoring. However, a seep has been coincidentally documented during marine habitat mapping east of the Otway Ranges (Ierodiaconou *et al.*, 2007). Video footage of this seep is provided in Appendix K. The source of the dissolved gas is either inputs from volcanic/magmatic activity or from decomposition of organic matter such as coal deposits. Several economic natural gas deposits, many dominated by CO_2 , have been discovered in the Otway Basin, whereas oil finds are much rarer (Mehin & Link, 1994). The natural gas deposits of the Otway Basin have been found to present certain magmatic characteristics (see Section 7.2.2). Therefore, in the discussion below, the isotopic composition of the CO_2 and H_2S in the basin is
assumed to be typically magmatic.

Establishment of a magmatic source for gases that are interacting with the groundwater in turn begs the question of mixing with magmatic water or H₂O vapour derived from magma degassing or dehydration. Most magmatic degassing releases gas which is dominated by H₂O, and in some cases may be more than 90% H₂O (e.g. Taran et al, 1995). Therefore, the the potential effects of addition of magmatic vapour H₂O or its condensate will be discussed alongside interaction with CO₂ and H₂S.

According to Faure (1986), water that has equilibrated with magma at temperatures greater than 700°C has an average composition ranging from 5.5 to 10‰VSMOW for δ^{18} O and from -50 to -85‰VSMOW for δ^{2} H. This range is very similar to that of Taylor (1997) for primary magmatic water, which is from 6 to 9‰ for δ^{18} O and from -40 to -80‰ for δ^{2} H and which was calculated via inverse modelling based on the isotopic composition of igneous minerals. Taylor's (1997) range will be used in this study. The isotopic composition of magmatic H₂O gas is related to that of the primary magmatic water, but the relationship is complex. In the case of a closed system and complete magma dehydration, all water is converted to H₂O vapour and hence the bulk composition of the two phases will be the same (Taylor, 1997). However, in an open system, significant partition of deuterium from the melt into the H₂O-vapour phase results in the exsolved H₂O_g having enriched δ^{2} H values. Accordingly, with continuing exsolution and escape of the vapour phase, the δ^{2} H value of the melt is continually depleted (e.g. Mark *et al.*, 2004). This, in turn, changes the composition of the vapour produced, ultimately resulting in increasingly depleted melt and vapour phases with the duration of exsolution (Taylor *et al.*, 1983).

An example of the magmatic water and water vapour fractionation system is shown in Figure 7.2, which also depicts the interaction between groundwater, magmatic water, CO_2 and H_2S . This diagram shows the compositional range of primary magmatic water, within which are plotted two hypothetical water compositions (A and B). Water A represents the composition of the H_2O in the melt before degassing of H_2O vapour which is isotopically heavier (e.g. compositions 1 or 2) and B represents the remaining, depleted H_2O in the melt. Fractionation from degassing is not well quantified but has been estimated as being between 10% (as in vapour #1) and 25% (as in vapour #2; Taylor, 1997).

If hydrogen and oxygen atoms exchanged between H₂O molecules of the magmatic vapour phase and H₂O molecules of the groundwater, there would be subsequent fractionation of both phases. The $\Delta_{(water-vapour)}$ at 100°C for deuterium is 24‰ and for oxygen is 5‰ (calculated from data from Clark & Fritz, 1997). The dark red arrows in Figure 7.2 show the result of this interaction at 100° given the isotopic compositions of the two phases, magmatic H₂O gas and groundwater. This process may occur to some very small extent at depth, however, it is much less likely to occur as the vapour rises through



Figure 7.2. Fractionation of groundwater molecules due to interaction with magmatic water or gases.

the hydrostratigraphic layers. This is because as it rises, it will cool and condense. Any change in isotopic composition arising from condensation can be ignored in this situation as all the vapour will condense. After condensation, the magmatically-derived H₂O will mix with the groundwater without fractionation. The result of said hypothetical mixing is depicted by dashed orange mixing lines in Figure 7.2. It is clear from this diagram that the process of interaction with both a magmatic H₂O vapour and/or its condensate results in offset of the groundwater composition to the right hand side of the LMWL. As this trend is not visible in the data from the Otway Basin groundwater, it is likely that only a minimal amount of magmatic H₂O, if any, has infiltrated the aquifers during the residence time of the *in situ* groundwater. This does not preclude the influence of magmatic CO₂ or H₂S on the groundwater and indicates that escape of water and gas from the plutonic or magmatic bodies has occurred at different stages and/or through different mechanisms.

The typical oxygen isotope composition of magmatic CO₂ (+10 ‰VMSOW) is shown on Figure 7.2 as a vertical green zone. The fractionation factor, (α), of ¹⁸O between CO_{2(g)} and water at 20°C is 1.04 (Clark & Fritz, 1997, p25). From Equation 7.2, Δ equals approximately 39‰. When CO₂ mixes with the groundwater, either as dissolved molecules or gas bubbles, the oxygen atoms of the CO₂ molecules exchange with those of the H₂O according to this fractionation factor, enriching the CO₂ and depleting the H₂O. Although magmatic CO₂ is enriched in heavy oxygen (i.e. the δ^{18} O of +10‰VSMOW) when compared to the source water of the Otway Basin (precipitation; e.g. -5‰VSMOW), mixing between the two results in further enrichment in the CO₂.

For example, a mixture of these components is considered where 1% of the oxygen atoms are sourced from the magmatic CO₂; this is a reasonable scenario given that systems have been observed with greater ratios (D'Amore & Panichi, 1985). With full equilibration, the exsolved CO₂ and the groundwater would have δ^{18} O values of 34.75 and -5.25 respectively. In this example, a change of 0.25% is observed in the groundwater, which is similar to the horizontal offset to the left of many samples in Figure 7.1. However, there are many other points that have greater x-axis offset; the maximum is 0.73%. Considering the full horizontal displacement from the LMWL as oxygen fractionation requires the assumption that no fractionation of the hydrogen isotopes has occurred, which is very unlikely. Further, if the full displacement were the result of CO₂ interaction, the mixing ratio needed to produce an offset of 0.73% would be much higher than 1%, which again, is not frequently observed in nature. In conclusion, it is most reasonable to believe that CO₂ interaction is not the only process which fractionates the groundwater.

To further constrain the dominance of CO_2 interaction on the groundwater fractionation, the ratio of mixing required to produce the x-axis offset observed was calculated. Several assumptions implicit in this estimate are: a) all the geogenic CO_2 encountered by the water undergoes full exchange of oxygen atoms, such that equilibrium is reached; b) temperature is 25° C; c) open system conditions allow for unlimited introduction and escape of CO₂; d) the full horizontal displacement from the LMWL is taken as oxygen fractionation; and e) geogenic CO₂ interaction is the main controlling factor that has caused fractionation of the oxygen in the water. The last assumption means that processes such as post-precipitation evaporation and dissolution of O-bearing molecules (e.g. CaCO₃ and soil CO₂) are not incorporated in the calculation.

The proportion of CO_2 interaction was estimated using a simple mass balance of oxygen atoms for the reaction:

$$CO_2^{16} + H_2O^{18} \to CO^{18}O^{16} + H_2O^{16}$$
 (7.7)

The mass balance incorporates the known isotopic values of the reactants (LHS) and the products (RHS) of the reaction in hypothetical proportions and is written as:

$$(\delta_{mag} \cdot 2a) + (\delta_{rech} \cdot (100 - 2a)) = (\delta_{DIC} \cdot 2a) + (\delta_{GW} \cdot (100 - 2a))$$
(7.8)

where δs are the δ^{18} O values of the appropriate index: mag is magnatic CO₂, rech is recharge water or precipitation, DIC is the exchanged CO₂ and GW is the depleted groundwater. The δ_{mag} value was assumed to be +10%; δ_{rech} was derived from the local meteoric water line; δ_{DIC} was calculated from the equation:

$$\delta_{DIC} - \delta_{GW} = \Delta = 40\% \tag{7.9}$$

and δ_{GW} was taken as the measured $\delta^{18}O$ values of the groundwater samples. The percentage of the total oxygen atoms that come from CO₂ was allowed to equal 2a. The remaining oxygen was attributed to the water, 100-2a.

Once the value of 2a is known, the molar ratio of CO_2 to H_2O can be calculated according to:

$$\frac{CO_2[mmol]}{H_2O[mmol]} = \frac{a}{100 - 2a}$$
(7.10)

and then converted to a molar:volume ratio (mmol:L) by using:

$$\frac{CO_2[mmol]}{H_2O[L]} = \frac{CO_2[mmol]}{H_2O[mmol]} \cdot k \tag{7.11}$$

where k is the number of millimoles in one litre of pure water (55,509). The results are plotted against the measured carbon concentration in the water samples in Figure 7.3, where it can be seen that the abundance of CO_2 deemed to have interacted with the water



Figure 7.3. The theoretical amount of magmatic CO_2 required to produce the fractionation of oxygen of groundwater samples is compared to the dissolved inorganic carbon (DIC) concentration, which was calculated using a charge balance of major and minor dissolved ions.

can be more than two orders of magnitude greater than the actual DIC. The differing values on the x and y axes of Figure 7.3 can be excused by the fact that not all CO_2 that interacts with the groundwater will be dissolved and converted to DIC. However, with an increase in the amount of CO_2 available for interaction, the partial pressure of CO_2 in the aquifer increases and leads to greater dissolution of the carbon in the water. Therefore, a positive relationship should be discernible between the two parameters. The fact that it is not leads to the conclusion that the assumptions for this calculation were false. Meaning that other processes, namely interaction with H₂S and silicate weathering are just as influential on the fractionation of the groundwater.

The expected $\delta^2 H$ composition of H₂S derived from the hydrogen of magmatic water at a temperature of 700°C is represented by a yellow shaded zone (Figure 7.2). This composition was calculated using fractionation data from Clark and Fritz (1997) under the assumption that the magmatic water has a $\delta^2 H$ range of -40 to -80‰.

At the time of mixing between H₂S and groundwater, the temperature is likely to be between 20°C and 70°C, which produces fractionation according to $1000ln\alpha$ values between 728 and 875, dictating that the $\Delta_{(water-H_2S)}$ value ranges from 829% to 1131% at 70°C to 20°C. These two fractionation factors were applied to a hypothetical situation where H_2S and groundwater are mixed at a ratio such that 0.5% of the total H atoms are sourced from H_2S and where δ^2H ranges of the groundwater and the gas were taken from Figure 7.2). The results show an enrichment of deuterium in the groundwater, a process indicated by yellow arrows in Figure 7.2. The magnitude of the increase in the groundwater δ^2H according to the hypothetical mixing ranges from 3.12 to 4.85%. Comparable vertical offset is visible in the difference between the groundwater and the LMWL in Figure 7.1.

Comparing the hypothetical mixing calculations for H_2S and CO_2 , it is clear that H_2S is more effective at fractionating the groundwater, even at smaller concentrations, because of the large fractionation factor involved. In reality, it is the combined effect of CO_2 and H_2S interaction and silicate hydration that has resulted in the displacement of the groundwater points from the local meteoric water line.

Interaction with soil CO_2 and aquifer carbonate

Low temperature water-rock interaction is a process capable of fractionating oxygen isotopes in groundwater such that samples plot to the left of the LMWL (inset in Figure 7.1). As interaction with silicates has been addressed above, this section is devoted to the effect of dissolution of carbonates and interaction with soil CO_2 in the Otway Basin.

While soil CO₂ is produced by root respiration in the subsurface, it is constantly exchanging with atmospheric CO₂, and therefore has an average δ^{18} O composition of $40\pm2\%$. Atmospheric CO₂ has this value because the ocean δ^{18} O value is 0% and the oxygen fractionation factor between them ($\Delta_{CO2-H2O}$ at 25°C) is 40% (Clark & Fritz, 1997). Therefore, mixing of soil CO₂ with a reservoir more depleted than the ocean (e.g. precipitation that is -5 - -4.5%) could produce no further enrichment of oxygen in the CO₂. On the contrary, one would expect the dissolved carbon to be slightly depleted when compared with soil CO₂ and the water to be slightly enriched after dissolution. Because a major enrichment of oxygen-18, independent of deuterium, is not observed in the groundwater when compared to precipitation, this fractionation is ruled out as unimportant in the oxygen system.

Dissolution of calcite from the aquifer matrix or cement contributes H₂CO₃ to solution; the reverse is possible via mineral precipitation. During these reactions, fractionation of the oxygen of the water molecules can occur. Alternatively, direct exchange of oxygen between water and the solid mineral phase can also occur. The fractionation is subject to an $\alpha_{(calcite-water)}$ value of 1.030 at 20°C. Evidence of fractionation of oxygen in calcite in the past via interaction with water within the basin has been demonstrated by Watson *et al.*, (2004). However, their work does not necessarily prove that carbonate interaction is currently an influential factor on the isotopic composition of the resident groundwater.

A calculation was performed to observe the effect on the Otway Basin groundwater

Temperature (°C)	Calcite δ^{18} O VSMOW	Calcite δ^{18} O VPDB	effect of interaction on groundwater $\delta^{18} {f O}$
20	greater than 25.8%	greater than -5.0%	enrichment
20	from 24.3% to 25.8%	from -6.4‰ to -5.0‰	little or no fractionation
20	less than 24.3%	less than -6.4 $\%$	depletion
80	greater than 15.5%	greater than -14.9%	enrichment
80	from 14% to 15.5%	from -16.4% to -14.9%	little or no fractionation
80	less than 14%	less than -16.4%	depletion

Table 7.1. The effect of various calcite compositions on groundwater fractionation.

of carbonate dissolution, precipitation or exchange. The temperature range used for the calculations was based on the range observed in the groundwater system today (approximately 20° C to 80° C). Carbonate interaction at the lower end of the temperature range is more pervasive in the Otway Basin than high-temperature interaction for two reasons. The first is that the deep, warm aquifers are mainly siliciclastic; whereas the majority of the carbonates occur in the upper units. The second is that elevated temperature conditions do not persist in much of the hydrogeologically significant extent of the basin. The terms, values and assumptions used in the calculations are: a) the groundwater δ^{18} O range is from -4% to -5.5%; b) the mineral is calcite, although in reality other carbonates are present, such as dolomite and ankerite; c) the fractionation factor, α , at 80°C is 1.020; and d) the fractionation factor, α , at 20°C is 1.030. The results of the calculations are summarised in Table 7.1. The calculations were performed using VSMOW values for calcite composition, but were converted to VPDB using an equation from Coplen et al. (1983) for comparison with published values relating to the composition of the basin's carbonates. Some cases in Table 7.1 result in depletion of heavy oxygen in the groundwater, which would place it to the left of the meteoric water line, as is seen in the current distribution of groundwater (Figure 7.1).

Whether or not interaction with calcite decreases the groundwater δ^{18} O value depends on the calcite composition. On the whole, the carbonates of the Otway Basin are greater than -5‰VPDB and hence will always cause enrichment of the groundwater (Table 7.1). Therefore, the groundwater trend to the left of the local meteoric water line cannot be explained by interaction with or dissolution/precipitation of carbonates. The specific isotopic compositions of some of the basin's aquifers are explored in more detail below.

Generally, marine calcite has an oxygen isotope composition dictated by fractionation from the contemporary composition of seawater and this is the case for Recent carbonates in the basin and the Upper Tertiary Carbonate Aquifer. The typical δ^{18} O composition of modern bryozoa from the Lacepede shelf, offshore from Robe, range from 0 to +2.0% VPDB (Bone & James, 1997). Ostracods from a lake near Coorong Lagoon in the Gambier Sub-basin have values of +1 to +4% VPDB (Edwards *et al.*, 2006). Miocene dolomite cement in the Upper Tertiary Carbonate Aquifer in the Port Campbell Sub-basin has δ^{18} O values of +0.1 to +0.6% VPDB (Nicolaides, 1997). Similarly, dolomite crystals of the the Upper Tertiary Carbonate Aquifer in the Gambier Sub-basin have values of -1.6 to +2.9% (Kyser *et al.*, 2002). Bulk rock composition of the Upper Tertiary Carbonate Aquifer in the Gambier Sub-basin has values of -4.2 to +1.4%, with the bryozoa and brachiopods of the Formation residing in the positive portion of this range (Kyser *et al.*, 1998). It is clear from the literature that the upper Tertiary carbonates of the Otway Basin have an δ^{18} O composition greater than -5% VPDB.

The Waarre Formation, in the lower part of the Upper Cretaceous Aquifer, has calcite and siderite cements with δ^{18} O values greater than 14‰VSMOW and less than 31‰VSMOW (Watson *et al.*, 2004). As conditions in this aquifer are mostly at elevated temperatures, the 80°C scenario (Table 7.1) may be applied, and either no fractionation or some enrichment of the groundwater may be expected from interaction with such carbonates.

No values of the oxygen isotope composition of the carbonate content of the Lower Tertiary Sandy Aquifer have been published to date. However, one may assume that, being partly primary marine cements like those of the Upper Tertiary Carbonate Aquifer, and partly secondary cements due to water-gas-rock interaction like those of the Upper Cretaceous Aquifer, the values of these carbonates are similar to those listed above.

The overall implication for the resident groundwater is that carbonate dissolution is currently not a major control on oxygen in groundwater. However, on a deeper consideration of the above values some past interaction is evident, and may be an important key to understanding the current system. For example, the limestone samples from the Upper Tertiary Carbonate Aquifer with the depleted compositions, those close to -4%VPDB, were noted by Kyser et al. (1998) to have higher proportions of secondary cements. This fact is also supported by James et al., 1993. The composition of these samples are close to the range of δ^{18} O values that are in equilibrium with groundwater at 20°C. Hence, the depletion in the secondary cements probably correlates to an enrichment of groundwater in the past. Indeed, these events occurred well in the past. Kyser et al. (2002) estimated the timing of the precipitation and dissolution events that produced the depleted dolomite cement in the Upper Tertiary Carbonate Aquifer, which involved at one point a mixture of meteroic-derived groundwater and sea water, to be before 1.6 Ma. Similar to the Upper Tertiary Carbonate Aquifer, some carbonate in the Upper Cretaceous Aquifer appears to be in equilibrium with groundwater and may therefore be considered to have formed via fractionation of the contemporaneous groundwater. However, this case is complicated

by the important role that magmatic CO_2 has played in the formation of these minerals (Watson *et al.*, 2004). In summary, precipitation and dissolution of carbonates in all the aquifers at any temperature may continue through time limited only by thermodynamics and kinetics. However, oxygen fractionation in the past may be limiting ongoing fractionation because a point near equilibrium may have been reached.

7.2.2 Carbon-13

Carbon-13 is the second most abundant isotope of carbon. The notation δ^{13} C compares the ratio of 13 C to 12 C in the sample to that of a standard. The data of this study are presented in ‰VPDB, which uses a Cretaceous marine fossil standard.

The main contributors of carbon to groundwater, and therefore the main controls on the fractionation of carbon isotopes, are introduction of soil CO_2 or magmatic CO_2 , dissolution of carbonate minerals, and production of CO_2 by subsurface biological reactions. These processes are addressed below with their relevance to the groundwater of the Otway Basin. In Section 7.2.1 the main reactions involving fractionation of oxygen were considered. Considerable overlap of these reactions exists with those listed above for carbon and therefore the findings are relevant to this discussion.

In total, 16 samples were taken for carbon isotope analysis. However, at two locations, the samples taken had too low a concentration of carbon for analysis and unfortunately resampling using a larger sample volume was not possible. A third sample was dismissed due to biological contamination. Finally, one last sample is discounted as it was taken as a repeat due to concern about contamination. As this sample returned very similar results to its double, it is not presented here (although all results are listed in Appendix F). Therefore, the final data set includes results for 12 boreholes, 10 of which represent the Lower Tertiary Sandy Aquifer and 2 of which are from the lower part of the Upper Tertiary Carbonate Aquifer (Clifton Formation). The values range from -7‰ to -22‰ (Figure 7.4).

The only sample from this study which lies in the Gambier Sub-basin is from the Dartmoor site (Figure 7.4). For comparison of groundwater from the east and west of the basin, some previously published data from the Lower Tertiary Sandy Aquifer are also incorporated in the map (Love *et al.*, 1994).

Soil CO_2

Atmospheric CO₂ generally has a carbon isotope composition of -6.4‰VPDB. After fractionation via photosynthesis and respiration of C3 plants, it is converted to soil CO₂ and has an average δ^{13} C value of -23‰ (Faure, 1986). Soil CO₂ has been measured in the



and two results from the Clifton Formation (green points at the Hotspur locations, this study). The data are underlain by the groundwater flow map for the Lower Tertiary Sandy Aquifer. Locations are shown in the inset, labelled by the closest town. Dartmoor, Heywood (NW), Portland, Port Fairy, Port Campbell Figure 7.4. A map of the δ^{13} C values in % VPDB of groundwater in the Lower Tertiary Sandy Aquifer (red points this study, blue points from Love *et al.*, 1994), and Carlisle River are town supply wells owned by Wannon Water. All other sites are part of the Victorian state observation network.



Figure 7.5. The variation in fractionation of carbon isotopes from soil CO_2 to DIC with pH due to the speciation of carbon at different pH ranges. Temperature variation is depicted by the different curves.

Gambier Sub-basin at a composition of $-22\% \pm 2\%$ VPDB (J. Dighton, unpublished data, 1984, as reported by Love *et al.*, 1994). Subsequent dissolution of soil CO₂ into groundwater is driven by, and proportional to, the partial pressure of CO₂ in the unsaturated zone. The process of dissolution, which entails formation of carbonic acid and dissociation to bicarbonate, also causes fractionation. The overall fractionation from soil CO₂ to total dissolved inorganic carbon (DIC) is determined by the fractionation factor between $CO_{2(soil)}$ and the three species of DIC possible depending on pH (refer to Figure 6.1). The variation in $\Delta_{(DIC-CO_2)}$ with both pH and temperature is illustrated in Figure 7.5.

The conditions in the recharge zones of the Otway Basin are variable, but are best exemplified by the western flanks of the Otway Ranges where the shallow groundwater pH is approximately 6.0 to 6.5, with a temperature of about 15°C to 18°C. Such conditions produce a fractionation ranging from 1.7‰ to 4.5‰ (Figure 7.5), resulting in a DIC composition between -20.3 and -17.5‰ (± 2). The majority of this range is due to the variation of pH, rather than that of temperature. Accordingly, groundwater DIC compositions measured in the recharge zone at Kawarren and Carlisle River are -20‰ and -22‰ respectively (Figure 7.6) and therefore plot close to the soil-derived DIC zone in Figure 7.6. This is good evidence of the dominant recharge process being rapid in the Port Campbell Sub-basin, as it appears from the results that little to no other carbon than soil CO₂ has been incorporated into the groundwater of these samples.

The third lowest value of δ^{13} C was -15‰, recorded in one of the Heywood bores,



Figure 7.6. Stable carbon isotope composition of groundwater DIC vs pH. Typical $\delta^{13}C$ values for various inputs to the Otway Basin are marked by grey areas; they are: soil CO₂ (-22‰ ± 2‰), volcanogenic CO₂ (-7‰ - -4‰) and Tertiary to Modern marine carbonate (not including secondary cement; 0‰ - +3.4‰). The dashed line represents DIC composition if derived only from soil CO₂ (-22‰) at 20°C (c.f. Figure 7.5). Samples are labelled by locations in Figure 7.4. KO: Koroit, PF: Port Fairy, HS: Hotspur, PC: Port Campbell, HW: Heywood, PO: Portland, PR: Princetown, CR: Carlisle River, KW: Kawarren & DM: Datmoor.

tapping the Lower Tertiary Sandy Aquifer. When compared to surrounding values, it seems anomalously low and may indicate mixing with modern recharge water from nearer the surface. Leakage is a likely candidate at this location as the structural integrity of the bore was visibly diminished at the time of sampling.

On the other hand, the recharge zone in the Gambier Sub-basin, as represented by the sample at Dartmoor and the northernmost samples from Love *et al.* (1994), does not exhibit carbon isotope compositions close to the juvenile signature. Instead, further fractionation or mixing of carbon isotopes has occurred in this region and resulted in values from -12.7% to -10.7%. This is a product of the recharge having interacted with carbonate rocks at or near the surface. This has a two-fold effect in changing the carbon-13 signature of the DIC. First, it results in dissolution of carbon with a marine signature (0 - -3.5%). Second, the carbonate-rich soil zone has higher pH than other areas. This in turn causes greater fractionation during dissolution of soil CO₂ (Figures 7.5 and 7.6).

Magmatic CO₂

Degassing of CO₂ from basaltic melt, although not entirely understood in every respect (Mattey, 1991), is a common enough geological process and it can contribute to natural gas stored in a basinal setting (e.g. Lollar *et al.*, 1997; Ballentine *et al.*, 2000). The carbon-13 signature of magmatically-derived CO₂ is dictated by the carbon composition in the melt and fractionation during degassing. Fractionation at high temperature is estimated to be between 1.8% and 2.2% and appears to be unaffected by reasonable variations in temperature and pressure (Mattey, 1991). As with degassing of water vapour and fractionation of deuterium, open versus closed system conditions produce different compositions of both components. Taking open system conditions into account, degassed CO₂ could have δ^{13} C values ranging from -3‰ to -5‰VPDB.

During exploration for petroleum in the Otway Basin, several instances of encountering CO_2 have been recorded. Wopfner and Thornton (1971) attributed the liquid CO_2 produced from wells Kalangadoo 1 and Caroline 1 to volcanic emanations, using the proximity of the then-called Newer Volcanics eruption centres (e.g. Mt Gambier) as a main source of evidence. Other sources of CO_2 were ruled out for this region due to the absence of hydrocarbons and carbonates in the source and trap horizons. Subsequent publications generally agreed on or confirmed a volcanic source of that CO_2 (e.g. Chivas *et al*, 1987; McKirdy & Chivas, 1992). Further evidence came from δ^{13} C analysis of CO₂ gas sourced from the basin and of the carbonate cements of the reservoir rocks and non-reservoir rocks (Watson *et al.*, 2004). Through this work, magmatic CO_2 was identified, although mixing with elements of crustal gas were not ruled out entirely. Watson et al. (2004) found that, where the percentage of gas was greater than 10% CO₂, the CO₂ of the Otway Basin bore high δ^{13} C values (-7 - -4‰), congruent with a mantle source (e.g. -5 - -3‰VPDB), rather than hydrocarbon association (-25 - -30%); Drever, 1997). In Waston's study (2004), as in that of Chivas et al. earlier (1987), helium isotopes were crucial in determining the source of the CO_2 .

In order to understand the effect on groundwater of introduced volcanogenic CO₂, it is necessary to consider the pH and temperature conditions during interaction. The former dictates the speciation of DIC and the latter controls fractionation. If the aquifer conditions throughout the basin are considered, other than those of the initial recharge stage, one may assume a minimum pH of 6.5 and a minimum temperature of 20°C. These conditions result in Δ_{DIC-CO_2} varying between 3‰ and 8.5‰ (Figure 7.5). If the CO_{2(volc)} has a composition of -7 - -4%, then its dissolution at the above rate of fractionation would produce DIC in the groundwater with a composition of -4 - +4.5% VPDB. Subsequent mixing of this DIC with that derived from soil CO₂ during recharge, which has a composition range of -20.3% to -17.5% (±2), would result in the total DIC having an intermediate δ^{13} C value. For example, a value of -11% is observed in several Otway Basin groundwater samples (Figure 7.6).

A similar discourse concerning the source of dissolved CO_2 in a body of groundwater and interaction of the two components can be seen in the literature pertaining to Hepburn Springs near Daylesford, Victoria (e.g. Lawrence, 1969; Shugg, 1996; Cartwright *et al*, 2002). At Daylesford it has been shown that volcanogenic CO_2 has interacted extensively with the groundwater, causing a depletion of oxygen-18 in the groundwater with respect to meteoric water and an enrichment of carbon-13 in the groundwater with respect to recharge DIC values (Cartwright *et al*, 2002). Carbon isotope values of DIC in groundwater from Hepburn Springs are -6–0‰VPDB (Cartwright *et al*, 2002).

Most of the carbon isotope compositions of the samples are enriched in carbon-13 (e.g. values of -11‰VPDB) with respect to initial recharge values of DIC (e.g. -20‰VPDB; Figure 7.4). This fractionation is in keeping with interaction with magmatic CO₂, however, it could also have been achieved via water-rock interaction with carbonates (see below). Fortunately, the effect on oxygen isotopes from these two processes has been shown to differ in the Otway Basin (Section 7.2.1). Therefore, by plotting the carbon isotope composition against the divergence from the LMWL in oxygen isotope composition (Figure 7.7), the two sources of carbon are discernible. Most of the data points plot to the left of the dotted line in Figure 7.7 and exhibit an enriched δ^{13} C value. Hence, they are likely to have been fractionated by interaction with magmatic CO₂.

As expected, the groundwater samples which were thought to contain only carbon from soil CO_2 , Kawarren and Carlisle River, do not differ greatly from the meteoric water line (Figure 7.7).

Dissolution of Carbonate Minerals

Dissolution of carbonate minerals from the aquifer matrix and/or cement is a source of carbon to the groundwater of the basin (Chapter 6). In order to gauge how significant a control over the carbon isotope composition this process is, the composition of the basin's carbonates must be known. The fractionation of carbon between calcite and bicarbonate ions is quite small ($\Delta_{calcite-HCO_3^-} = 0.7\%$ at 20°). Therefore, reactions between these two phases where they initially have differing compositions will result in the homogenisation of both.

The majority of the basin's carbonates are contained within the modern sediment and the upper Tertiary units, although significant amounts of carbonate cement exists in the older formations also.

The typical carbon isotope composition of modern bryozoa from the Lacepede shelf,



Figure 7.7. The x-axis shows the divergence from the LMWL of the groundwater samples as represented by the difference in their δ^{18} O values at the same δ^{2} H value. Positive values represent offset to the right of the LMWL and negative to the left. The divergence is plotted against the δ^{13} C values of the groundwater samples. The analytical precision of δ^{18} O and δ^{13} C are $\pm 0.15\%$ and $\pm 0.2\%$ respectively. Sample locations and grey input areas are explained in Figure 7.6.

offshore from Robe, is -0.2–2.4‰VPDB (Bone & James, 1997). Miocene dolomite cement in the Upper Tertiary Carbonate Aquifer from the Port Campbell Sub-basin has values of 2.5–2.6‰VPDB (Nicolaides, 1997) and the same mineral in the Gambier Sub-basin is 0.7– 2.8‰ (Kyser *et al.*, 2002). The bulk rock composition of the Upper Tertiary Carbonate Aquifer from the Gambier Sub-basin ranges from -9.5 to 2.1‰ (Kyser *et al.*, 1998).

There is no published data pertaining to the carbon isotope ratios of the carbonaceous matter contained in the Lower Tertiary Sandy Aquifer. Although, Watson *et al.* studied the deeper aquifer system (the Waarre Formation of the Upper Cretaceous Aquifer) and found the carbonate cements therein to have values ranging from -15 to 5% VPDB (2004).

The isotopic composition of marine carbonate has been fairly constant over the Cretaceous and Tertiary and hence δ^{13} C values are close to that of the international standard, the PeeDee Belemnite, which is 0‰VPDB. For a more refined base-level composition of marine carbonates in the Otway Basin, one may take the composition of Miocene brachiopods from the Balcombe Clay in the nearby, contemporaneous Port Phillip Basin. The brachiopods have values ranging from 3.0 to 3.4‰VPDB (Nicolaides, 1997). The marine carbonates that produce bulk rock values significantly lower than this (e.g. the Upper Tertiary Carbonate Aquifer from the Gambier Sub-basin; Kyser *et al.*, 1998) most likely contain diagenetic cement which has altered the isotopic content.

Freshwater carbonates generally have much lower δ^{13} C values, e.g. -15‰ (Faure, 1986), as seen in the Upper Cretaceous Aquifer carbonate. Carbonate values that deviate significantly from their original composition are likely to be the result of fractionation from interaction with CO₂ from various sources, e.g. CO₂ from organic oxidation of methane (Faure, 1986) or volcanogenic CO₂. Indeed, this is the case for the carbonate cement of the Upper Cretaceous Aquifer that has higher δ^{13} C values. They are in fact late stage cements and are enriched in carbon-13 due to interaction with CO₂ with a composition of about -5‰ (Watson *et al.*, 2004).

Dissolution of carbonate of the Upper Tertiary Carbonate Aquifer during diffuse recharge to the Lower Tertiary Sandy Aquifer is expected as the water in the latter is generally saturated with respect to calcite, especially in the Gambier Sub-basin (Love *et al.*, 1994). Given that the carbonate compositions (0-3.4%) are generally greater than the values expected from DIC derived from soil CO₂ (-20.3 – -17.5\%), enrichment in carbon-13 of the DIC is to be expected. Most groundwater samples yielded DIC enriched in carbon-13 when compared to recharge water (Figure 7.7). However, few points plot to the right of the dotted line in Figure 7.7, meaning that those samples do not exhibit lasting evidence of the fractionation of oxygen istopes of the water caused by dissolution of carbonates. One sample that shows a strong carbonate interaction signature is from the Upper Tertiary Carbonate Aquifer (Clifton Formation), which fits well with the carbonate-rich lithology of the aquifer.

It could be concluded from these results that carbonate dissolution has not occurred in the groundwater. However, this is not the case, because other lines of evidence have shown that the process occurs. Rather, it is clearer to state that evidence of carbonate weathering is not visible in the oxygen isotope results and the weathering effect on the carbon isotope signature cannot be distinguished from that produced via interaction with volcanic CO_2 .

Biogenic CO_2

Johns (1968) was possibly the first author to document the high alkalinity of the Otway Basin groundwater, noting that it well exceeded the expected HCO₃ concentration based on normal soil CO₂ inputs from recharge zones and that alkalinity varied largely independently of chlorinity. He attributed the excess HCO₃ to production of CO₂ by bacteria in the subsurface and decomposition of carbonaceous matter from the aquifer matrix, as these are common inputs in other areas. However, given the total lack of strongly positive δ^{13} C values in the groundwater samples, and the strongly established source of volcanogenic CO_2 in the basin, biogenic inputs are unlikely to be major.

Carbon dioxide and calcite dissolution

When looking at the spatial distribution of the δ^{13} C values (Figure 7.4), it is clear that progressive enrichment along-flow does not occur. There are depleted values in the eastern recharge zone, however, enrichment does not simply increase with position in the flow regime. This is due to the addition of volcanogenic CO₂ via faults and fractures, or other, perhaps more diffuse, paths. This addition is not spatially indiscriminate; it is reliant upon the existence of delivery pathways. However, it does account for the seemingly random array of slightly different δ^{13} C values along the flow paths of the Lower Tertiary Sandy Aquifer. Yet, it is not just the addition of CO₂ that drives the enrichment. The added acidity causes decreased pH and therefore increased dissolution of carbonates from the aquifer matrix and continued dissolution in waters close to saturation with calcite. Therefore, an amplifying mechanism exists in the carbon system whereby the dissolution of both CO₂ and carbonate minerals are connected and both enrich ¹³C in the DIC.

Volcanogenic carbon dioxide has been suggested by others in the past to be a driver for carbonate dissolution in the Otway Basin , with substantive evidence relating to the formation of cenotes within the Upper Tertiary Carbonate Aquifer (Lewis, 1984) and stromatolites within the cenotes (Webb *et al.*, 2000). It is proposed that the preferential pathways of upwards-travelling CO_2 causes upwards migration of dissolution of limestone and forms cenotes and may have enhanced phreatic cave formation (White, 2005). Evidence of the cenotes having been formed via bottom-up dissolution includes their extensive depths. The true bottom of the cenotes is often obscured by collapsed limestone material; in other cases the cenote breaches the entire extent of the limestone formation (Lewis, 1984). Figure 7.8 shows the location of the cenotes and other karst features in relation to volcanic eruption centres of the basin. In addition to the morphological evidence offered by the cenotes, other caves proffer information on the migration of CO_2 within the system. For example, several caves in the Upper Tertiary Carbonate Aquifer have elevated concentrations of carbon dioxide in their atmospheres and one in particular (Mt. Burr Cave) has lethal levels of the gas (S. White, pers. comm., 2008).

Love *et al* (1994) did not credit the enriched δ^{13} C values of Gambier Sub-basin groundwater to any input from volcanogenic CO₂ and concluded that, in regions where the vertical hydraulic gradient was neutral or upwards, the similar δ^{13} C values and the calcite saturation index meant that there were no reactions occurring to fractionate the carbon isotopes of the groundwater from the Lower Tertiary Sandy Aquifer. Similarly, Bennetts (2005), concluded that calcite saturation indices in the lower part of the Upper Tertiary Carbonate Aquifer near Condah precluded the possibility of ongoing (or even past) car-





bonate dissolution. Both cases have significant repercussions in terms of the application of a correction for carbonate dissolution in carbon-14 dating. Furthermore, the suggestion that future karstification at saturated depths is impossible (Love *et al*, 1993) may be incorrect.

7.2.3 Strontium-87

Strontium-86 and strontium-87 are the second and third most abundant stable isotopes of strontium respectively. Non-radiogenic strontium-87 coexists with that produced by the radioactive decay of rubidium-87, present in minerals. Strontium-87 concentration is conventionally represented as a standardised ratio: ${}^{87}\text{Sr}/{}^{86}\text{Sr}$; ratios in rocks and minerals may vary systematically depending on mineral type and/or rock age. The oceanic ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ value is stable at any one time due to effective mixing on a shorter time scale than that of the geological cycle of strontium through the oceanic reservoir (Hess *et al*, 1986).

Strontium isotope ratios in groundwater are a product of interaction processes, be it mixing, dissolution or ion exchange, with various sources of strontium, namely: rocks, rainfall and dry fall, e.g. continental dust (McNutt, 2000). It also is known that biological cycling of strontium can effect the strontium isotope composition (Graustein & Armstrong, 1983). Rocks with organic matter contain strontium with 87 Sr/ 86 Sr ratios that were inherited from the organic matter prior to its deposition. The strontium ratio of organic matter grew before it was deposited (Gosz & Moore, 1989; Blum *et al.*, 2008; Garcia-Ruiz *et al.*, 2008). In turn, the strontium ratio of soil and atmospheric dust in any one environment is dictated by strontium sources, such as rocks and seawater, and their relative proximity (Gosz & Moore, 1989; Gingele *et al.*, 2007; Negrel *et al.*, 2007; Raiber *et al.*, 2009). These sources and processes are discussed in relation to the groundwater of the Otway Basin below.

Modern rainwater at the coast generally has a strontium ratio similar to that of the ocean (0.70917; Herut *et al.*, 1993). However, rainwater composition tends to change with distance from the coast due to loss of marine salts and addition of pollution or dust from the continent. Often these latter components are more radiogenic and, therefore, ratios of precipitation have been observed in several studies to be higher inland (e.g. Andersson & Lofvendahl, 1990; Aubert *et al.*, 2002; Raiber *et al.*, 2009). Indeed, this is the case in the Otway Basin, although rainfall values there are still very close to the marine ratio (Raiber *et al.*, 2009). In contrast, some regions have inland rainfall ratios that are lower than 0.70917 (e.g. Herut *et al.*, 1993; Han & Liu, 2006). These cases are usually in regions where the surface geology has a major component of marine carbonate or may also be

affected by anthropogenic inputs.

The variation of the ocean's strontium composition is on a time scale too long to effect the composition of rainfall that may have become water which is now hosted within the Otway Basin aquifers. More specifically, the ratio of the ocean has only varied between approximately 0.70917 and 0.7091 in the last million years (Hess *et al.*, 1986). However, the low sea level of 20 ka resulted in higher continentality of the basin's recharge zones in the past and also the exposure of currently-submerged surface geology. These changes in the Otway Basin region may mean that rainwater from periods of the Quaternary had slightly different strontium isotope ratios, either higher or lower than modern rain. The difficulty lies in quantifying the variation and, although beyond the scope of this study, may be a product of future work.

Strontium ratios in marine sediment, both carbonate and non-carbonate, are a product of the authigenic components, which will have the same ratio as the ocean at the time of deposition, and allogenic components, which are largely either marine volcanics or siliceous terrigenous material which, as such, have inherited Sr ratios. Volcanic inputs have an average ratio of 0.704 ± 0.002 (Faure, 1986) and of course may constitute terrigenous material in some cases. Siliceous allogenic material has an average strontium ratio of 0.720 ± 0.005 (Faure, 1986). The strontium ratio in the ocean has changed markedly over geological time and has been documented by the study of the above-mentioned authigenic components of marine sediment. A history of change was published by Hess et al. (1986); some key values for the Otway Basin and some global averages, are depicted in Figure 7.9. Athough the plotted marine carbonate ratios are only those of Hess et al. (1986) and James et al. (1993), they are generally consistent with other samples from the Otway Basin, namely, those of Kyser et al (1998) and Webb et al. (2000). Interaction with groundwater and/or other diagenetic fluid can have an effect on the strontium isotope composition of the aquifer matrix (James et al., 1993). The Eocene carbonate composition depicted in Figure 7.9 represents primary marine cement and clasts only (Hess *et al.*, 1986).

The dissolution of calcium-bearing minerals is especially important in influencing the strontium isotope ratio of groundwater because strontium often substitutes for calcium due to the similarities between these two elements. It is this effect that causes relatively high strontium concentrations (dominated by ⁸⁸Sr and ⁸⁶Sr) in carbonates. However, due to the paucity of the parent isotope, ⁸⁷Rb, ⁸⁷Sr concentrations and ⁸⁷Sr/⁸⁶Sr ratios are usually low in carbonates (Fisher & Stueber, 1976).

Conversely, silicates can be a source of ⁸⁷Rb and therefore have comparatively high ratios (Figure 7.9). This is important in the Otwayn Basin because the aquifer matrix of the Lower Tertiary Sandy aquifer is composed not only of the carbonate constituent, which is mostly in the form of primary and secondary cements, but also of the quartz-



Figure 7.9. Strontium isotope ratio values for rocks and groundwater of the Otway Basin (this study; references b, f & g) as compared to some global averages (references c, d & e). Superscripts denote references. The vertical grey zone represents rainfall values of samples from Casterton, Hamilton & Heywood (reference g). Ranges labelled as an aquifer or an aquitard represent water samples from that host (number of samples in brackets); all other ranges, except sea water, represent bulk rock values.

dominated clastic grains. Because of their largely terrigenous origin, these grains would have strontium isotope ratios in the range of 0.715–0.725 (Figure 7.9).

The exception to silicates having high ratios is young volcanic rock, which usually has even lower values of ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ than carbonates. For example, young basalt of Mt Cameroon, western Africa, has an average strontium isotope ratio of 0.703325 (Benedetti *et al.*, 2003). The composition range of the basalt of the Pliocene-Quaternary Volcanic Aquifer is marked in Figure 7.9 (Price *et al.*, 1997).

The results of the strontium isotope analysis of 34 groundwater samples from the various hydrostratigraphic units of the Otway Basin are depicted in Figure 7.9, where they can be compared to the relevant influences mentioned above. The Otway Basin groundwater samples have 87 Sr/ 86 Sr values that range from the bulk rock composition of the Pliocene-Quaternary Volcanic Aquifer to that of local rainfall; this range also overlaps that of Tertiary marine carbonates. Therefore, these three components are likely to be influential on the groundwater ratios.

However, no samples from the aquifers are more enriched in ⁸⁷Sr than rainwater for the region, meaning that siliceous terrigenous material is not influential (Figure 7.9). This is in contrast to the groundwater samples of Raiber *et al.*, which are from a neighbouring region in similar aquifers (2009; Figure 7.9). Their samples contain evidence of significant dissolution of silicates from nearby outcrops of Palaeozoic granite bedrock and from the deep lead sediment (sourced in part from the granite; Raiber *et al.*, 2009). The Otway Basin has some basement rocks of granitic composition, e.g. the Padthaway Ridge, but not in abundance (Chapter 2). The main geological units in contact with the groundwater sampled in the Otway Basin that contain abundant silicate minerals are the Eumeralla Formation (Lower Cretaceous Basement) and the Lower Tertiary Sandy Aquifer. Therefore, the lack of groundwater ratios in the Otway Basin greater than the ratio of rainfall indicates that the degree of weathering of silicate minerals from these formations is relatively low compared to weathering of the volcanic or carbonate rocks.

When the strontium ratio is plotted against the sample depth (Figure 7.10) the influences on ⁸⁷Sr/⁸⁶Sr become more easily discernible. Some samples from the Plio-Pleistocene Heterogeneous the Upper Tertiary Carbonate Aquifers are strongly influenced by the volcanic aquifer composition, with ratios significantly lower than late Tertiary carbonate rock (Figure 7.10). This confirms the importance of the recharge pathways through volcanic eruption centres which have been identified through the localised highs in hydraulic head coinciding with variability in the salinity (Chapters 4 and 6). Additionally, dissolution of the marine carbonate has undoubtedly affected the strontium ratio of the groundwater of the Upper Tertiary Carbonate Aquifer, as all samples have lower values than rainfall.

The effect of carbonate dissolution is also visible in the Lower Tertiary Sandy Aquifer in



Figure 7.10. Stable strontium isotope ratio variation in all groundwater samples with depth of the bottom of the sreen interval of the bores sampled. Vertical shaded areas represent rainwater and bulk rock strontium compositions taken from sources listed in Figure 7.9. $HW^{\#2}$ is the leaky bore at Heywood. With the exception of this bore the samples from the Lower Tertiary Sandy Aquifer form a trend marked by the dashed lines.

Figure 7.10, where it is shown that the ratio in this groundwater decreases with increasing sample depth (the exception being the sample from the leaky bore at Heywood). This decline illustrates both the interaction with the younger carbonates during recharge to the deeper aquifer and ongoing incorporation of strontium from Eocene carbonate in the aquifer itself. The latter process highlights the solubility of the aquifer's carbonate content over its quartz and silicate content; and it also supports the conclusion that addition of volcanic CO_2 creates a shift in pH that drives carbonate dissolution even in the deep, confined levels of the Aquifer. Although, even if the water becomes saturated with respect to calcite, strontium exchange can still affect the isotopic signature.

To further constrain this idea, the strontium isotope signatures are plotted in Figure 7.11 against the available δ^{13} C values from Figure 7.11. These samples are all from the Clifton Formation or the Lower Tertiary Sandy Aquifer. The carbon isotope composition of the samples of Carlisle River and Kawarren indicate that the majority of carbon is from



Figure 7.11. Stable strontium isotope ratio variation with $\delta^{13}C$ values for groundwater samples. Typical $\delta^{13}C$ values of soil CO₂, volcanogenic CO₂ and marine carbonate are marked by horizontal grey areas. Vertical shaded areas represent rainwater and bulk rock strontium compositions. Samples are labelled by locations in Figure 7.4. KO: Koroit, PF: Port Fairy, HS: Hotspur, PC: Port Campbell, HW: Heywood, PO: Portland, PR: Princetown, CR: Carlisle River, KW: Kawarren & DM: Datmoor.

the soil zone and very little water rock interaction has occurred at this early stage of the flow system. Despite this, the strontium ratio for these samples is lower than recorded rainfall composition and sits in the range of young carbonate rocks. Interpretation of these results leads to the conclusion that the 87 Sr/ 86 Sr ratio of rainfall in the area of the western foothills of the Otway Ranges may be closer to that of current ocean water than the rainfall depicted in Figure 7.11 (i.e. that of rainfall measured at Heywood, Hamilton and Casterton; Raiber *et al.* 2009; Figure 7.9). This fits well with the location of the young groundwater samples being closer to the coast than the rainfall sites. Those samples that do fall within the measured rainfall range, Heywood #2 and Dartmoor, confirm the influence of recent recharge (Dartmoor) or leakage of juvenile water (Heywood #2) on their composition.

It is interesting to note that the groundwater sampled at Dartmoor plots very close to the strontium rainfall range but does not have a δ^{13} C value close to that of soil CO₂. This would indicate that a process has fractionated the carbon in the water during or since recharge, without changing the strontium content a great deal. Given a soil water pH of 7, a similar pH to that observed in the water at Dartmoor, soil CO₂ fractionates to DIC with a composition of -15.35 % VPDB. The main candidate for further fractionation to the Dartmoor composition (-10.9% VPDB) is dissolution of volcanogenic gas.

The main cluster of sample points in Figure 7.11 show clear evidence of the dissolution of marine carbonates, both of the overlying Upper Tertiary aquifer and aquitard and, as evidenced by the water from Port Fairy and Koroit, of the Lower Tertiary Sandy Aquifer matrix itself. The reason this dissolution is distinct from interaction with the volcanic rocks, which also have low strontium ratios, is that the decrease in the strontium ratio is accompanied by an increase in $\delta^{13}C$. Hence, both CO₂ gas and carbonate mineral interaction can be identified in the water of the Lower Tertiary Sandy Aquifer.

7.3 Radioactive Isotope Hydrogeochemistry

7.3.1 Radiocarbon

Introduction

Carbon-14 is a cosmogenic isotope. It is formed in the reaction:

$$n + {}^{14}N = {}^{14}C + H \tag{7.12}$$

where the bombardment of nitrogen atoms by neutrons from cosmic rays results in transmutation of the nitrogen atom. Carbon-14 is radioactive; it decays to nitrogen via beta emission according to:

$${}^{14}C = {}^{14}N + e^- + \bar{v}_e \tag{7.13}$$

at a constant rate, which is defined by the decay constant, $\lambda = 0.000121$ years⁻¹ (Clark & Fritz, 1997). Radiocarbon in the atmosphere is transferred to soil CO₂ by vegetation and then dissolved by recharging groundwater as DIC; this part of the carbon cycle forms the basis of groundwater radiocarbon dating.

The cosmic ray interaction with our atmosphere and the subsequent steady decay of radiocarbon is responsible for a relatively constant activity of radiocarbon in the atmosphere at any one time which is then transferred to DIC in groundwater. This quantity provides the input of A_0 , or initial atmospheric activity, to Equation 6.1 for water age calculation. The activity in the late 1800s was 13.65 dpm/g and this value is taken to be 100%mC (percent modern carbon), the value used in the age calculation in Equation 6.1. However, A_0 has varied considerably in geological time due to changes in the intensity of cosmic radiation penetrating the earth's atmosphere. The variation is known from measurements of radiocarbon in trees and corals which have been dated using independent methods (e.g. Bard *et al.*, 1993). Variation in A_0 should be accounted for in radiocarbon dating of groundwater by way of a calibration, although this process is often omitted. In recent years, much progress has been made in creating calibration systems for converting radiocarbon years to calendar years (e.g. special issue: *Radiocarbon*, 2004, **46**, [3]). Unfortunately, it was not within the scope of this study to account for past atmospheric activity being greater or less than 100%mC, so results recorded in this study remain in radiocarbon years.

Further complicating the A_0 issue is the fact that radiocarbon can be produced due to neutron availability through sources other than cosmic rays. Neutrons can be produced in the subsurface during decay of radioactive elements such as Uranium. The radionuclides produced via spallation from these subsurface neutrons are termed hypogenic. As the Otway Basin does not have concentrated radioactive mineral deposits, hypogenic ¹⁴C is discounted in this study. Neutrons were also produced in proliferation by nuclear weapons testing in the middle of the 20th century. The resultant increase in radionuclide activity in the atmosphere is termed the "bomb-pulse" and was detected globally. Any groundwater that was recharged after this time has been affected and exhibits %mC values greater than 100. Exchange, mixing or diffusion from bomb radiocarbon with water containing older carbon is an issue in unconfined areas and recharge zones; such conditions can yield values in samples less than 100%mC and are therefore more difficult to interpret.

A limitation of radiocarbon dating of groundwater is that by measuring carbon decay, one is assessing a solute of the groundwater, rather than the water particles themselves. A simple model of carbon decay involves the assumption that carbon-14 is a conservative solute and will travel at the velocity of the groundwater without loss or gain of mass, other than that from decay. Unfortunately, this assumption is known to be false. Addition of radiocarbon may occur along flow due to hypogenic inputs or via mixing or exchange with younger water bodies. Apparent loss of radiocarbon can occur along flow due to its analytical measurement being relative to total carbon. For example, with addition of DIC which has a ¹⁴C activity of 0%mC the overall activity of ¹⁴C in the groundwater is effectively lowered. Such dilution is partially accounted for by the application of a correction (see below). In recent years the importance of dilution of ¹⁴C in aquifers via leakage of older groundwater from adjacent aquitards has been brought to the attention of the research community (Goode, 1996; Bethke & Johnson, 2002). The issue has been most effectively tackled by reactive transport modelling and the expression of groundwater age in a new parameter of age mass (Bethke & Johnson, 2008). In this study, aquitard leakage is only taken into account in so far as the dilution correction applied below allows.

Dilution Corrections

Before interpreting the radiocarbon ages calculated using Equation 6.1, a correction for the dilution due to dead carbon is necessary. In order to determine the most appropriate correction formulae, the processes that are likely to be contributing the carbon to the groundwater must be known. The sources of carbon to the groundwater of the basin have been explored using isotope geochemistry in Sections 7.2.1, 7.2.2 and 7.2.3. It has been shown that, aside from introduction of soil CO_2 during recharge, carbon is incorporated into the DIC of the groundwater from both carbonate mineral dissolution and interaction with volcanogenic CO_2 .

The DIC incorporated into groundwater from soil CO_2 is the component that once represented A_0 in the groundwater and hence is the component needed for measuring decay since recharge. Dilution is caused by addition to groundwater DIC of carbon which does not contain ¹⁴C, or contains very little. In order for carbonate minerals to contain undetectable amounts of ¹⁴C, they must be older than 60,000 years, or about ten ¹⁴C half-lives. More specifically, consider a mineral which contained 100%mC at the time of precipitation, 60,000 years ago. Observed now, after burial causing isolation from the atmosphere and ocean, decay has resulted in less than 0.1%mC persisting in the mineral, a virtually undetectable amount (Freeze & Cherry, 1979). All the aquifer material of the Otway Basin except parts of the Plio-Pleistocene Heterogeneous Aquifer and the Pliocene-Quaternary Volcanic Aquifer are older than 60,000 years and therefore will contain negligible carbon-14. Additionally, magmatic CO_2 is generated from carbon originally derived from old limestone, therefore, it usually contains no ¹⁴C (Mook, 2006). Thus, both processes have a diluting effect on groundwater ¹⁴C.

Equation 6.1 was used to calculate the raw radiocarbon age of the groundwater (results are recorded under t_{RAW} in Table 7.3). In order to apply corrections, one simply adds a correction factor, q, to the equation such that it becomes:

$$t = \frac{\ln(\frac{A_t}{A_0 \cdot q})}{-\lambda} \tag{7.14}$$

Where t is time (radiocarbon years), A_0 is the ¹⁴C activity of the DIC at the time of recharge (100%mC), A_t is the ¹⁴C activity of the DIC at the time of sampling (%mC), and q is a constant ranging from 0 to 1, derived by various correction methods.

Four different q values were derived and were applied to the results to give a range of corrected radiocarbon ages of the groundwater samples. The models are described below and the inputs and results are recorded in Tables 7.2 and 7.3 respectively. Essentially, the q value can be regarded as the proportion by which the initial carbon-14 activity (A_0) has been diluted by the dissolution of dead carbon. For example, if the q value is 0.85 it

implies that dissolution of the dead carbon has resulted in the dilution of A_0 to 85% of its original concentration.

The first q value used for corrections was q_{ALK} which follows the methodology of Clark and Fritz (1997). The model uses the concentration of carbonic acid and bicarbonate in the groundwater sample to estimate the amount of carbonate dissolution that has occurred. The equation used was:

$$q_{ALK} = \frac{m_{H_2CO_3} + 0.5m_{HCO_3^-}}{m_{H_2CO_3} + m_{HCO_2^-}}$$
(7.15)

where the m is the molar concentration of the species denoted in subscript.

The model for q_{ALK} assumes that dead carbon dissolution occurs in closed system conditions, and that no exchange with soil CO₂ may occur after recharge (Clark & Fritz, 1997). This model is potentially a good fit for the aquifers sampled, considering they are confined. However, if any addition of volcanogenic CO₂ or leakage and mixing were to occur in the subsurface, the model becomes invalid. Further, another problem is that, because carbonic acid concentration becomes insignificant in comparison to bicarbonate concentration if the pH exceeds 7, q_{ALK} is limited to a value of 0.5 at that pH range. In Figure 7.12 the q_{ALK} correction factors for each sample are plotted against the δ^{13} C value for that sample. One point with a q_{ALK} value of 0.5 can be seen to have the highest δ^{13} C value. The latter indicates significant carbon dissolution that is not accounted for by the former; this is a reasonable assertion considering that the q values by other methods for that sample are lower (Figure 7.12).

The remaining correction models were derived from the $\delta^{13}C$ values measured in the groundwater samples (denoted as $\delta^{13}C_{DIC}$). The first of these is the Gonfiantini model, (Gonfiantini, 1972; as cited by Geyh, 2000). The correction factor is $q_{\delta^{13}Ccarb}$, which is designed to account for dissolution of dead carbon from carbonate minerals in the soil zone and the aquifer matrix. It is defined as:

$$q_{\delta^{13}Ccarb} = \frac{\delta^{13}C_{DIC} - \delta^{13}C_{carb}}{\delta^{13}C_{rech} - \delta^{13}C_{carb}}$$
(7.16)

where $\delta^{13}C_{carb}$ represents the isotopic composition of dissolved carbon derived from dissolution of carbonate minerals. This value is assumed to be 0‰ due to the fact that most of the carbonate minerals in the basin have a similar value. There is some fractionation that takes place during the dissolution of calcite at the temperature range considered, however, this is disregarded as relatively insignificant (Section 7.2.2). The term $\delta^{13}C_{rech}$ is the $\delta^{13}C$ value of DIC derived from dissolution of biogenic soil CO₂ during recharge and is defined as:

Bore ID or supply town	Code	Aquifer	Screen* (m)	$\mathop{\rm A}_t (\% {\rm mC})$	error (%mC)	DIG Der Q	CO ₂ (mmol/L)	HCO_3^- (mmol/L) (DIC mg/L)	pH_{rech}	Δ_{rech}	$\delta^{13} \mathrm{C}_{rec}$	${}^{h} \operatorname{T}_{volc}$ (°C)	pH_{volc}	Δ_{volc}	$\delta^{13} C_{volc}$
146024	$\mathrm{HS}^{\#1}$	UTC	153	<2.0	NA	-12.6	0.00	4.06	154	6.5	4.4	-17.6	20	7.5	7.8	1.8
146023	$\mathrm{HS}^{\#2}$	UTC	98	6.1	± 0.1	2-	0.00	6.04	154	6.5	4.4	-17.6	20	7	6.6	0.6
Port Campbell	PC	\mathbf{LTS}	009	8.7	\pm	-11	2.88	4.57	405	6.0	1.7	-20.3	30	6.5	3.8	-2.2
97995	\mathbf{PR}	LTS	471	27.1	± 0.1	-12	0.00	1.88	125	6.0	1.7	-20.3	15	6	9.0	3.0
Carlisle River	CR	LTS	148	70.6	± 0.3	-22	9.77	1.02	492	5.5	0.0	-22.0	15	5.5	0.0	-6.0
108910	КW	\mathbf{STJ}	140	63.9	± 0.3	-20	0.36	1.26	93	6.0	1.7	-20.3	15	6.5	4.4	-1.6
325468	$HW^{#2}$	\mathbf{STJ}	613	37.5	± 1.1	-15	0.00	1.64	254	6.5	4.4	-17.6	20	10	8.0	2.0
Portland	РО	STJ	1000	3.1	± 0.9	-11	0.16	11.68	716	6.5	4.4	-17.6	40	7.5	5.9	-0.1
Heywood	$HW^{\#1}$	STJ	500	3.9	± 0.9	-11.8	0.33	6.90	436	6.5	4.4	-17.6	30	7.5	6.8	0.8
Port Fairy	ΡF	STJ	850	$<\!2.0$	NA	-11	0.40	6.02	385	6.5	4.4	-17.6	40	7.5	5.9	-0.1
72659	КО	STJ	1499	$<\!2.0$	NA	-10.1	0.91	8.43	554	6.5	4.4	-17.6	40	7.5	5.9	-0.1
Dartmoor	DM	TTS	500	82.6	± 0.3	-12	0.32	5.55	352	7.0	6.9	-15.1	20	7	6.6	0.6

Table 7.2. Inputs for ¹⁴C dilution corrections: %mC, δ^{13} C, alkalinity, pH, temperature and fractionation difference (Δ) values.

Bore ID or supply town	Code	A_t (%mC)	q_{ALK}	$q_{\delta 13}C carb$	$q_{\delta 13}Cvolc$	$q_{\delta 13C} carb+volc$	t_{RAW}		t_{ALK}	$t_{\delta 13Ccarb}$	$t_{\delta 13Cvolc}$	$t_{\delta 13Ccarb+volc}$
146024	HS#1	<2.0	0.50	0.72	0.74	0.53	>32,400	$\pm 3,000$	>26,600	>29,600	>29,900	>27,200
146023	$\mathrm{HS}^{\#2}$	6.1	0.50	0.40	0.42	0.17	23,100	土 430	17,400	15,500	16,000	8,400
Port Campbell	PC	8.7	0.69	0.54	0.49	0.26	20,200	\pm 1,010	17,200	15,100	14,200	9,100
97995	\mathbf{PR}	27.14	0.50	0.59	0.64	0.38	10,800	主 700	5,100	6,400	7,100	2,800
Carlisle River	CR	70.6	0.95	1.00	1.00	1.00	2,880	± 35	2,480	2,870	2,870	2,870
108910	КW	63.9	0.61	0.99	0.98	0.97	3,700	± 40	-370	3,580	3,570	3,450
325468	$\mathrm{HW}^{\#2}$	37.5	0.50	0.85	0.87	0.74	8,110	± 250	2,380	6,810	6,950	5,660
Portland	РО	3.1	0.51	0.63	0.62	0.39	28,700	$\pm 2,840$	23,100	24,900	24,800	21,000
Heywood	$HW^{\#1}$	3.9	0.52	0.67	0.69	0.46	26,800	$\pm 2,170$	21,500	23,500	23,700	20,400
Port Fairy	ΡF	$<\!2.0$	0.53	0.63	0.62	0.39	>32,400	\pm 3,000	>27,100	> 28,500	> 28,500	>24,600
72659	КО	$<\!2.0$	0.55	0.58	0.57	0.33	>32,400	\pm 3,000	>27,400	>27,800	> 27,700	>23,200
Dartmoor	DM	82.6	0.53	0.80	0.81	0.64	1,580	\pm 85	-3,720	-290	-210	-2,080
Note: All ages (Note: The error	(t values) a s for the u	are in radioca	rbon year	S.	from the %	in Seular values in	Tahle 7.9 ar	also annly	the comected :	Sebe		
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Figure 7.12. The four different ¹⁴C dilution correction factors (q values) plotted against δ^{13} C values for the dated groundwater samples.

$$\delta^{13}C_{rech} = \delta^{13}C_{soil\ CO_2} + \Delta_{rech} \tag{7.17}$$

where $\delta^{13}C_{soil\ CO_2}$ is taken as -22‰ and Δ_{rech} is the fractionation difference determined by the fractionation from soil CO₂ to dissolved carbon at the recharge pH and temperature conditions (Section 7.2.2). The temperature conditions were assumed to be 15°C everywhere and the pH conditions were judged based on the pH of the water sample itself or the pH of shallow water in its regional recharge zone. Recharge pH varies throughout the basin, as already discussed; the values used for each sample are listed as pH_{rech} in Table 7.2, along with the Δ_{rech} and $\delta^{13}C_{rech}$ values. The $q_{\delta^{13}Ccarb}$ values derived from these calculations are shown in Table 7.3.

The weakness of this correction is in the assumed value for $\delta^{13}C_{carb}$ (0‰). The marine carbonate in the aquifer matrix has variable δ^{13} C values due to differing primary composition and due to diagenetic processes or exchange with groundwater since deposition (Section 7.2.2). In addition, non-marine carbonate exists at the surface which has different values again. This uncertainty is not uncommon in other studies, but a value of zero is conventionally used (Geyh, 2000). The third correction factor is also based on the Gonfiantini model, however, the dead carbon input of carbonate mineral dissolution is replaced by the input of dissolved geogenic or magmatic CO_2 . The resultant correction factor is described thus:

$$q_{\delta^{13}Cvolc} = \frac{\delta^{13}C_{DIC} - \delta^{13}C_{volc}}{\delta^{13}C_{rech} - \delta^{13}C_{volc}}$$
(7.18)

where $\delta^{13}C_{DIC}$ and $\delta^{13}C_{rech}$ have previously been defined and $\delta^{13}C_{volc}$ describes the isotopic composition of the portion of DIC derived from the inorganic CO₂. It is calculated by:

$$\delta^{13}C_{volc} = \delta^{13}C_{volc\ CO_2} + \Delta_{volc} \tag{7.19}$$

where $\delta^{13}C_{volc\ CO_2}$ is assumed to be -6‰ and Δ_{volc} is the fractionation difference determined by the fractionation from volcanogenic CO₂ to dissolved carbon at the *in situ* pH and temperature conditions. The generalised temperature and pH conditions were judged based on those of the water sample itself, taking into account the large screen depth or long flow paths in some cases. The prescribed values for the dated samples are listed as T_{volc} and pH_{volc} respectively in Table 7.2 with the resultant $\delta^{13}C_{volc}$ compositions. The correction factor values, $q_{\delta^{13}Cvolc}$, are listed in Table 7.3.

The fourth dead carbon dissolution factor was designed to account for both the dissolution of carbonate minerals and inorganic CO_2 . It is the product of the second and third corrections factors, and is expressed as:

$$q_{\delta^{13}Ccarb+volc} = q_{\delta^{13}Ccarb} \times q_{\delta^{13}Cvolc} \tag{7.20}$$

This style of combining two δ^{13} C correction factors is suggested by Clark and Fritz (1997) when two sources of dead carbon are considered to be important in the one system. However, according to Mook (2006), there is no simple method to distinguish the effect of gas and carbonate sources using the δ^{13} C values when there is a mixture of the processes in the water samples due to the similarity of the carbon isotope signatures of the inputs. Indeed, the two dead carbon sources being considered in this study have very similar isotopic compositions in terms of their contribution to DIC, demonstrated by the fact that all but two of the values of $\delta^{13}C_{volc}$ in Table 7.2 are within 2.3% of the $\delta^{13}C_{carb}$ value of 0%.

In order to further explore the relationship described in Equation 7.20, an hypothetical mixing model with three carbon components was constructed with varying proportions of recharge DIC, ranging from 10% to 100%. The hypothetical inputs used were recharge DIC with a δ^{13} C value of -22%, carbon from mineral dissolution of 0.5% and dissolved



Figure 7.13. Comparison of dilution correction factors using hypothetical mixing of three carbon components: recharge/soil DIC, carbonate minerals and volcanogenic CO_2 .

carbon from volcanogenic CO₂ of 1‰. The proportions of recharge carbon used for each iteration of the mixing calculations were compared to the proportions expected from the dilution fractions, which were calculated using the final mixure composition, along with the mixing components' compositions. The results are presented in Figure 7.13, which shows the difference between the actual proportion of recharge carbon input to the mixing calculation (dotted line, x axis) and the predicted proportions derived from the dilution models described above (y axis). The results clearly show that in a case such as this, where the two dead carbon inputs are such similar compositions, the values of either $q_{\delta^{13}Ccarb}$ or $q_{\delta^{13}Cvolc}$ provide a much better approximation of the component of original carbon from soil CO₂ than the $q_{\delta^{13}Ccarb+volc}$ value does. Therefore, the fourth dilution correction model is dismissed as over-correcting in such cases. The results from this model are not utilised in the interpretation of the data for this study (although they are recorded in Table 7.3 under $t_{\delta^{13}Ccarb+volc}$).

As the q_{ALK} and $q_{\delta^{13}Ccarb+volc}$ factors have been dismissed, the validity of the remaining factors must be confirmed such that they can be applied to the data. The results of the hypothetical mixing model show that the two remaining q factors are effective at estimating the actual proportion of dilution (Figure 7.13). Although the two carbon sources have different compositions, namely 0‰ and -6‰, the differing rates of fractionation of calcite and CO₂ upon their dissolution in water means that the dilution factors, $q_{\delta^{13}Ccarb}$ or $q_{\delta^{13}Cvolc}$, are very similar (Figure 7.12. The fact that they are very similar to each other and different to the q_{ALK} and $q_{\delta^{13}Ccarb+volc}$ values is an important indicator of the real amount of dilution of the recharge carbon. It is with this evidence that the use of corrected ages t_{carb} and t_{volc} in the interpretation of the radiocarbon results is justified. Rather than using both ages, one or the other, or an average, the youngest corrected age of the two values for each sample will be used in the interpretation. This ensures that the scenario for maximum dilution is taken into account in each case.

Discussion

The corrected results of ¹⁴C analyses [radiocarbon years] at 13 sites, 11 sites for the Lower Tertiary Sandy Aquifer and 2 for the deeper part of the Upper Tertiary Carbonate Aquifer (the Clifton Formation), are presented in Figure 7.14. The samples from the Clifton Formation were taken at the Hotspur locations. Two sites are marked in red due to the unreliable nature of their results. The first unreliable sample was from a state observation bore near Heywood, yielding an uncorrected age of 8,100 years. As mentioned previously, at the time of sampling it was clear this bore was nearing the end of its life and it was apparent that some failure of the screen or casing may have occurred, allowing leakage from the surface or other aquifers. When compared to the different result from the nearby Heywood town supply bore, which is in very good condition, it is confirmed that this sample represents a mixture of water from the Lower Tertiary Sandy Aquifer and younger water from downwards leakage. Other parameters, e.g. ⁸⁷Sr/⁸⁶Sr, also provide further proof that this bore is contaminated by shallower water. The second failed location is Nirranda, where the groundwater sample taken was not voluminous enough to provide adequate dissolved carbon for the analysis and resampling was not possible.

At three points in Figure 7.14, Hotspur, Koroit and Port Fairy, values are given as being greater than the corrected age (e.g. >29,600), as these samples contained less than 2%mC. The ages on the map where derived from correcting the raw age of 32,400 radiocarbon years, which corresponds to an A_t value of 2%mC. When ¹⁴C activity is as low as 2%mC, it may be due to the water being old, i.e. almost all the original radiocarbon has decayed, or it could be the result of addition of dead carbon to the water or mixing with water of an even older age from, perhaps, an aquitard. Regardless of the reason, such low activity levels flag the approach of the limit of the application of this dating method and the limit of its analytical precision. For the purposes of better understanding the flow system using all available data, the corrected ages derived from a value of 2%mC are quoted in the discussion, although it is important to recognise that those ages are minima.



Figure 7.14. A map of the corrected radiocarbon ages (green), ³⁶Cl R values (black) and ³⁶Cl A values (blue) of groundwater in the Lower Tertiary Sandy Aquifer, with two results from the Clifton Formation (Hotspur locations). The error ranges for the ³⁶ Cl R and ³⁶ Cl A results are 1.6–2.4 and 1–14 respectively. Unreliable radiocarbon sites are labelled in red. Locations are shown in the inset, blue locations are for ³⁶Cl rainfall data. Kawarren, Koroit, and Hotspur are flowing artesian wells; all three, along with Princetown, are part of the state network. All other sites are town supply wells owned by Wannon Water, except Port Macdonnell, whose radiocarbon age is from Love *et al.* (1994); see Table 7.2 for details.

The groundwater ages are young, less than 4,000 radiocarbon years, in recharge zones near Lower Tertiary Sandy Aquifer outcrop, i.e. the Dartmoor site in the Gambier Subbasin and the Carlisle River and Kawarren sites in the Port Campbell Sub-basin. Contrastingly, the age of the groundwater in the Clifton Formation at the inland Hotspur sites is much older. Coastal sites are generally further along the flow path and they also exhibit old ages. Although ages are presented for at least each sub-basin, the sampling was restricted to suitable bores of reliable condition, of which there are few in Victoria. The result is that the piston flow model cannot be applied as it was for the cross sections of the study by Love *et al.* in South Australia (1994). However, in the Tyrendarra Sub-basin, decay of ¹⁴C is apparent in the age increase of 3,530 radiocarbon years is visible between Carlisle River and Princetown. These observations, while useful, should not be interpreted as accurate and actual residence times for the groundwater because of the inherent problematic closed system assumptions of the piston flow model (Bethke & Johnson, 2008).

The age of the water at Kawarren is older than than at Carlisle River, despite Kawarren sitting higher in the regional flow system (Figure 7.14). This can be explained by the nature of the flow system in that region (Section 4.8). Essentially, the southwestwards flow from Kawarren continues south through connection with the groundwater of the Upper Cretaceous Aquifer and rejoins the groundwater of the Lower Tertiary Sandy Aquifer again near Carlisle River. The reason the groundwater does not get gain age along this path is because significant recharge and subsequent mixing of younger water occurs in the Upper Cretaceous Aquifer, which is exposed at the surface there.

Both the Kawarren and Carlisle River groundwater ages are significantly older than that of the water at Dartmoor. The reason for the older ages in the recharge zone of the Port Campbell Sub-basin is not simply longer flow paths or slower flow, but rather, a combination of these factors with a lack of very modern leakage. Contribution of very modern water containing bomb-pulse ¹⁴C is the most likely cause of the negative age of groundwater at Dartmoor in the Gambier Sub-basin. Therefore, in these cases, the Upper Tertiary Aquitard is more effective at preventing leakage in the east than it is near Dartmoor.

Additionally, the rapid infiltration of very young water at Dartmoor is likely to be aided by karstic secondary porosity in the overlying Upper Tertiary Carbonate Aquifer (Figure 7.8). Secondary porosity has a direct affect on the age of water, where an aquifer with higher fracture or karst feature density correlates to groundwater with younger ages (e.g. Hadzisehovic *et al.*, 1993). This can be due to actual groundwater flow being faster in fractures and karst conduits than in primary porosity, and is certainly the case in the
Gambier Sub-basin (Herczeg *et al.*, 1997). However, groundwater radiocarbon age can become artificially older due to exchange of carbon with the aquifer matrix. Loss of ¹⁴C from solution via exchange with matrix material is greater in water within the intergranular pore space than within fractures (Geyh, 2000). Hence, some of the difference in the age of water from secondary and primary porosity systems is real and some is apparent. Some authors incorporate this effect by considering radiocarbon ages as maxima (Love *et al.*, 1994), whereas others have developed a relationship between the actual and apparent ages dictated by the ratio of primary porosity to fracture porosity (Maloszewski & Zuber, 1984). In the case of Dartmoor, flow aided by transport through karst has been less affected by exchange and, therefore, the young age is effectively maintained.

The groundwater of the Clifton Formation, sampled near Hotspur, is dated as 15,500 and 29,600 years, which is significantly older than might be expected, given that it is from a shallower aquifer and is significantly up-gradient from Heywood, where the age is 23,500 radiocarbon years. Additionally, the large difference in the ages from the two boreholes, which are only 7.6 km apart cannot be explained by aging along the flow path because one bore does not sit directly down-gradient of the other (Figure 7.14; Bennetts, 2005). It is proposed here that the reason for these two differing and old ages is leakage of older water from the Upper Tertiary Aquitard into the Clifton Formation, induced by the significant groundwater extraction from the latter for irrigation (Chapter 5). The Clifton Formation is very susceptible to the age effect of this leakage in the Condah region because it is sandwiched between thick sequences of the Gellibrand Marl and the Narrawaturk Marl of the Upper Tertiary Aquitard (Figures 2.4 and 6.11). The relative thicknesses of the hydrostratigraphic units is is important as it influences the ratio of their groundwater volumes, which is the factor that affects the amount of age change due to leakage (Bethke & Johnson, 2002).

Lastly, the radiocarbon ages produced for the Lower Tertiary Sandy Aquifer groundwater add weight to the conclusions made in Section 6.3.5 about the control of past climatic fluctuations on the salinity of recharge water. For example, groundwater at the points of Heywood and Koroit is quite fresh (EC is 1,500 and 2500 μ S/cm respectively), but lies down-gradient of more saline water (Figure 6.16). The radiocarbon ages of the water at these locations (23,500 and 27,700 radiocarbon years respectively) indicate that the groundwater is probably derived from precipitation that fell during the pluvial climate phase which prevailed prior to the aridity of the last glacial maximum around 20 ka (Chapter 3). The fact that both old water (e.g. at Portland) and young water (e.g. at Carlisle River) are fresh proves that the accumulation of dissolved solids is not due solely to water-rock interaction and the alternative explanation of changes to the P/E ratio is necessary.

7.3.2 Chlorine-36

Chlorine-36 is a radioactive isotope with a half-life of $301,000 \pm 4,000$ years (Phillips, 2000) which is produced by spallation of ³⁶Ar, ⁴⁰Ar, ³⁵Cl, ³⁹K or ⁴⁰Ca, and which decays to ³⁶Ar by beta particle emission or, more rarely, to ³⁶S via electron capture (Clark and Fritz, 1997). There are three main sources of the chlorine-36 found in groundwater; they are: cosmogenic, epigenic and hypogenic. Cosmogenic production occurs in the atmosphere where the above-mentioned spallation is driven by cosmic ray interaction with argon gas. Epigenic production occurs at the earth's surface as a result of cosmic ray interaction with earth minerals and salts. Hypogenic production, as with hypogenic ¹⁴C production, occurs in the subsurface and is not related to cosmic rays. Hypogenic ³⁶Cl is produced via reaction between stable atoms (e.g. ³⁵Cl) and neutrons sourced from radioactive decay of U, Th, or K within the aquifer matrix. A bomb-pulse of ³⁶Cl was created by addition of neutrons from island- or ship-based nuclear weapons testing, which activated oceanic ³⁵Cl (Bentley et al., 1982). The rate of fallout reached a peak around 1955–1960 of 1000 times the background level, but had dissipated completely, returning to the pre-bomb level, by 1964 (Bentley et al, 1982). The bomb-pulse peak in ³⁶Cl did not persist as atmospheric fallout, despite the long halflife of the isotope, due to effective washout, meaning that chloride ions in the atmosphere readily form salts and/or are dissolved in water droplets and are delivered to the surface as wet or dry precipitation.

The measured abundance of 36 Cl in groundwater may be reported as an atomic ratio, R:

$$R^{36}Cl = \frac{number \ of \ atoms \ of \ ^{36}Cl}{number \ of \ atoms \ of \ all \ Cl}$$
(7.21)

or as an atomic concentration in solution, A:

$$A^{36}Cl = \frac{number \ of \ atoms \ of \ ^{36}Cl}{litres \ of \ solution} = R^{36} \cdot mCl \cdot N_A \tag{7.22}$$

R can be converted to A by using Equation 7.22, where mCl is the concentration of chloride (as moles/L) and N_A is the Avogadro constant.

The measured R and A values for groundwater samples can be used to calculate the time elapsed since recharge, t, using the equation (Phillips, 2000):

$$t = \frac{-1}{\lambda} \cdot \ln \frac{C_t (R_t - R_{se})}{C_0 (R_0 - R_{se})}$$
(7.23)

where R_t and C_t are the measured ³⁶Cl/Cl ratio and Cl⁻ concentration respectively, and R_0 and C_0 are the same parameters for the simulated initial recharge water; R_0 takes into account cosmogenic and epigenic sources of ³⁶Cl. R_{se} is the ³⁶Cl/Cl ratio that represents a hypothetical secular equilibrium, the balance between constant hypogenic production and radioactive decay. Estimation of R_{se} should not take into account any comsogenic or epigenic inputs. This equation assumes the cosmogenic, epigenic and hypogenic inputs are constant; they can be estimated or measured. The estimations made for this study regarding these inputs are outlined below.

Because of the long halflife of the isotope, decay over a timeframe of 30,000 years cannot be detected by changes in the isotope's activity (Bird *et al.*, 1989) and it is usually suggested that the limit of its use is much older, e.g. 100,000 years (Mazor, 1992) or 500,000 years (Phillips, 2000). As carbon dating has shown that some of the groundwater from the Otway Basin is certainly younger than 30,000 years, the application of Equation 7.23 was made to the data with the anticipation that the method may not be viable.

Additionally, the history of the chloride in the groundwater is important when using it to understand the history of the water (Davis *et al.*, 1998a), as the two are not one and the same. Notwithstanding this, the relatively conservative behaviour of chloride in the Otway Basin system established in Chapter 6 lends the system well to the use of this isotope as an environmental tracer.

Estimation of ³⁶Cl in precipitation

Lal and Peters (1967) produced a latitude-dependent equation for calculating chlorine-36 fallout. Bentley *et al.* (1986) point out that continentality is an important factor in ³⁶Cl fallout due to the addition of chloride from sea spray close to the coast. Davie *et al.* (1989b) calculated the expected R value of precipitation over south-east Australia using Lal and Peters' formula (1967) and detailed chloride concentrations (Hutton & Leslie, 1958). The results showed increased R values with distance from the coast and therefore illustrated that the formula simulates continentality through the inclusion of chloride concentration. In addition, a later study made direct measurements of ³⁶Cl fallout in Australia (Keywood *et al.*, 1998) and confirmed the applicability of Lal and Peters' formula (1967) to the southern hemisphere, with a minor alteration.

Keywood *et al.* (1998) measured ³⁶Cl fallout to the earth's surface along two transects inland from the coast in Australia and showed that fallout is signifantly less in the southern hemisphere than in the northern hemisphere. They found the average fallout for the locations, apart from some anomalously high monsoon measurements, was 15.4 atoms m^2/s .

An original estimation of modern R_0 has been made for the recharge zones of the Lower Tertiary Sandy Aquifer in the Otway Basin using this published data. Rainfall volumes with chloride concentrations for 4 locations (Hutton & Leslie, 1958) were used in conjunction with the average fallout of 15.4 atoms m²/s (Keywood *et al.*, 1998). Hence, the following values for $R_0 \times 10^{15}$ were calculated: 18 (Cressy), 20 (Derrinallum), 11 (Cavendish) and 10 (Coleraine)(Figure 7.14). Herczeg *et al.* (1997) reported that a bulk winter rain composition for the northern Gambier Sub-basin (at Kybybolite) had an R value of 15×10^{-15} , whereas summer rainfall was enriched in ³⁶Cl, at a value of 60×10^{-15} . According to the calculations of Davie *et al.* (1989b), modern precipitation over the Otway Basin should have an $R \times 10^{15}$ value equal to or less than 20. The estimated values comply with this and in addition are comparable to a rainfall value of 20 in the Mallee region of north-west Victoria (Bird *et al.*, 1989; 1991). Most of these values are significantly lower than those associated with bomb-pulse ³⁶Cl, which is in the order of 40 to 350 (Bird *et al.*, 1991).

Davis *et al.* (1998a) gave a comprehensive overview of R_0 estimation methodologies. A factor identified as crucial in any ³⁶Cl study is the variation in atmospheric production over the last several tens of thousands of years, which is depicted in Figure 7.15. For most of the recent past, the production has been higher than today. Specifically, there has been a declining trend approximately from the time of the last glacial maximum to today. The repercussion of this is that R_0 at the time of past recharge is likely to have been higher than modern R_0 . Therefore, the modern Otway Basin values ranging from 20 in the east to 10 in the west may act as minimum estimates of $R_0 \times 10^{15}$. Additionally, an hypothetical palaeorecharge R_0 was devised with a value of 29.9, reflecting the greater cosmic ray activity and a changed chloride concentration in rainfall because of lower sea level in the past.

\mathbf{R}_{se} estimation

In Section 7.3.1 hypogenic production of radiocarbon was dismissed in the Otway Basin due to the rarity of radioactive minerals. However, the same assumption for hypogenic ³⁶Cl is not necessarily a straightforward one, as sometimes hypogenic production of ³⁶Cl can be greater than cosmogenic fallout and bomb-pulse inputs (Andrews *et al.*, 1989). A method for calculating expected R_{se} values for rocks from their neutron flux rates has been developed and utilised in several studies (Andrews *et al.*, 1989; Table 7.4). Some specific published values are a presented in Table 7.4; they were chosen for their representative spread and for the types of host rocks studied, such that the values could be useful as estimates of the R_{se} of the hydrostratigraphic units of the Otway Basin.

The dominant lithologies of the the Lower Tertiary Sandy Aquifer are interbedded sand and clay or mud, for which the sandstone and shale listed in Table 7.4 is analogous. Shale $R_{se} \times 10^{15}$ may range from 5 to 14, whereas sand is generally less radioactive, at 2 to 10. The carbonate value from Table 7.4 is 9 and indicates that the Clifton Formation is likely to have a R_{se} value to that of the confined aquifer. Basement rocks have much higher



Figure 7.15. Relative change in production of 36 Cl in the atmosphere over recent geological time, after Davis *et al.*, 1998a.

values (Table 7.4) but the groundwater exposure to their rock matrix is limited compared to the long travel times in the aquifers themselves. Therefore, to perform groundwater age calculations for these aquifers of the Otway Basin, a range of 2–14 seems appropriate for $R_{se} \times 10^{15}$.

However, the relationships of the parameters in the calculation are limited by the logarithm in Equation 7.23, such that:

$$\frac{C_t(R_t - R_{se})}{C_0(R_0 - R_{se})} > 0 \tag{7.24}$$

Therefore, R_{se} must be either less than both of R_0 and R_t , or greater than both, for the age calculation to be viable. As discussed above, R_0 has been estimated as ranging from 10 to 29.9 and the measurements of R_t from the nine groundwater samples are 8.5–15.6 (Figure 7.14). Therefore, rather than a range of 2–14, the R_{se} is prescribed as 2–8.4.

Interpretation of the ³⁶Cl results

Groundwater from the Lower Tertiary Sandy Aquifer was sampled at 8 locations, and one sample was taken from the Clifton Formation, for analysis of dissolved ³⁶Cl. The results

Location or Forma- tion	Ref.	Rock type	U (mg/kg)	Th (mg/kg)	Cl (mg/kg)	Neutron flux $cm^{-2}yr^{-1}$	$^{36}\text{Cl/Cl} \times 10^{15}$
Saratoga Springs	1	carbonate	2.2	1.7	150	631	9
Fruitland Coal	2	coal				270	4
Saratoga Springs	1	basement	3.5	18	150	2020	29
Äspö Diorite, Sweden	3	diorite					49
Saratoga Springs	1	shale	3.2	11	150	947	14
Great Artesian Basin	4	shale				915	13
Milk River Aquifer	5	shale				331	5
Saratoga Springs	1	siliciclastic	0.45	1.7	150	145	2
Great Artesian Basin	4	sandstone				410	6
Milk River Aquifer	5	sandstone				674	10

Table 7.4. R_{se} values (³⁶Cl/Cl) of various rock types. References: 1—Rao *et al.*, 2005; 2—Snyder & Fabryka-Martin, 2007; 3—Louvat *et al.*, 1999; 4—Lehmann *et al.*, 2003; 5—Andrews *et al.*, 1991.

were provided in the map in Figure 7.14 as ${\rm ^{36}Cl/Cl}$ ratios (R) and ${\rm ^{36}Cl}$ concentrations (A).

The age calculation (Equation 7.23) was performed for each groundwater sample, despite the expectation that only those samples that were too old to be dated by ¹⁴C (i.e. Port Fairy and Koroit; Figure 7.14) could possibly produce viable results. For each sample the range of R_0 and R_{se} values were used to determine the sensitivity of the age to these inputs. The results were mostly negative ages, confirming that the groundwater is generally too young to be dated by the long-lived ³⁶Cl. There were two samples, those with some of the lowest ³⁶Cl ratios (i.e. Dartmoor and Kawarren; Figure 7.14), that produced positive ages. Given that these are two of the three youngest samples according to the radiocarbon dating and that they are located in recharge zones, the positive ages are false. They only occurred when the highest R_0 values were employed in the equation, which artificially simulated decay from the initial ratio (high) to the current activity ratio (low). In reality, it is more likely that the lower R_0 values are appropriate for these sites, and therefore, the t values are negative and the water is too young to be dated.

This result from the dating is not a defeat of the application of 36 Cl to this groundwater because the isotope results can still provide much more information about the history of the water and its solutes. First, statements regarding the age of the water may still be made using the 36 Cl data. Second, the negative ages mean that radioactive decay of 36 Cl is not the main control of the abundance of the isotope. Hence, the question of which processes are influential is the next avenue of investigation. Third, the 36 Cl/Cl ratios are useful in determining the source of chloride in the groundwater (Andrews *et al.*, 1989).

The failure of the dating provides an indication that the water sampled in this study

was most likely recharged before about 100,000 ka, otherwise some decay of ³⁶Cl should have been detected. This means the old water at Port Fairy and Koroit that is close to the limit of radiocarbon dating is from the intermittently pluvial period of the basin's history that occurred after the last interglacial (c. 130 ka), but before the last glacial maximum (c. 20 ka; Chapter 3). A further conclusion is that all the samples were recharged more than 50 years ago as no bomb-pulse ³⁶Cl was detected. All the ³⁶Cl/Cl values for the water from the Otway Basin are less than 15.7×10^{-15} and the youngest water (e.g. Dartmoor and Carlisle River) has the lowest ratios. For comparison, ³⁶Cl/Cl ratios expected from water recharged after 1950 are more than 40×10^{-15} and may be up to 350×10^{-15} (Bird et al., 1991). From Section 7.3.1, some water containg bomb-pulse ¹⁴C is likely to have mixed with the groundwater at Dartmoor, this effect is not visible in the chloride data, maybe because the signal is too diluted.

Numerous processes that affect the chloride-36 concentration and/or the ³⁶Cl/Cl ratio of groundwater aside from decay have been identified and graphical methods have been shown to be useful in understanding spatial trends (e.g. Bird *et al.*, 1989; Davie *et al.*, 1989b; Bird *et al.*, 1991; Mazor, 1992). Figure 7.16 shows the groundwater ³⁶Cl/Cl ratios plotted against the ³⁶Cl concentrations. Significant decay of the isotope along flow results in a concurrent decrease in both parameters (inset of Figure 7.16). This change is observed at one point of the flow system, namely from Heywood to Portland (Figures 7.14 and 7.16). However, it is concluded that this is an apparent trend of decay only, because of the negative t values calculated for the samples and because the difference in their ¹⁴C ages is 1,300 radiocarbon years. Contrastingly, the opposite relationship is observed from Carlisle River to Port Campbell (Figure 7.16), which may indicate the production or addition of ³⁶Cl. Because of these conflicting results, it is clear that neither ³⁶Cl decay nor hypogenic production are strong influences on the ³⁶Cl distribution in solution. The young age of the water precludes decay and the low R_{se} values of the aquifer matrix compared the the ratios in the water mean that an increase in R_t is not due to hypogenic production.

When the 36 Cl/Cl ratios are plotted against the corrected radiocarbon ages, a generally positive trend is discernible (Figure 7.17), meaning that the older samples contain more radioactive chlorine. Obviously, this trend cannot be due to the progressive addition of dead Cl⁻ with a ratio close to zero, e.g. old connate salt. As stated, hypogenic production is unlikely given the anticipated low neutron flux of the sandy aquifer material. However, it is possible that diffusion of chloride with a different 36 Cl/Cl ratio can occur from adjacent aquitards into an aquifer (Fabryka-Martin *et al.*, 1991). This process is possible given the geology of the Otway Basin, with significant aquitard layers within and adjacent to the Lower Tertiary Sandy Aquifer. For the process to cause an increase in 36 Cl/Cl along the flow path of the aquifer, the ratio must be higher in the aquitard. A reason this might



Figure 7.16. The ³⁶Cl concentrations (A) and the ³⁶Cl/Cl ratios (R) of groundwater samples from the study. The inset shows processes that affect the parameters, after Davie *et al.*, 1989b. Two arrows are shown for the dataset, drawn from one point to another which is almost directly down-gradient (c.f. Figure 7.14).

happen is the slightly higher neutron flux rates of shale when compared to sand (Table 7.4).

However, the conclusion of this study is that the groundwater still retains a 36 Cl/Cl ratio close to its original value. The differing ratios with age are then interpreted to be due to the fluctuation of 36 Cl/Cl in rainfall in the past. It has been established that the atmospheric production of 36 Cl has changed over time, (Davis *et al.*, 1998a; Figure 7.15). When the relative productivity is plotted on the same time scale as the groundwater samples for this study (Figure 7.17), a strong coincidence of fluctuations on the respective x-axes is visible. In particular, this theory explains well why both increases and decreases in 36 Cl/Cl are seen along flow paths (e.g. Figure 7.16). The two main outliers are the Port Fairy and Koroit samples that contained less than 2% modern carbon and hence were probably too old to be dated with 14 C. If this connection between the measured 36 Cl/Cl ratio in the groundwater and the 36 Cl/Cl value of the rainfall of the past from



Figure 7.17. R³⁶Cl values of groundwater samples plotted against corrected ¹⁴C ages and compared to the 1000-year moving average of the relative production rate of atmospheric ³⁶Cl in the last 40,000 years (Davis et al 1998a; same as Figure 7.15).

which it is derived is real, these two samples may be as old as 37,000 years. An increase in 36 Cl/Cl along the flow path has also been observed in the Mallee region of the Murray Basin, north of the Otway Basin. Similar to the conclusion of this thesis, Bird *et al.* (1989) attributed the increase at this location to variability in R₀ of rainfall, rather than hypogenic production.

Lastly, the source of the chloride in the groundwater can be constrained using the 36 Cl/Cl measurements. As stated above, the increase in the 36 Cl/Cl ratio with the age of the groundwater proves that the progressive addition of dead chloride (36 Cl/Cl = 0) is not possible. This then precludes the idea that a significant source of the salt in the water is from insufficiently-flushed connate water, because the proposed connate water in the formations would contain negligible 36 Cl (Davie *et al*, 1989b). This finding supports the conclusion of this study, namely that the elevated concentration of dissolved solids in some parts of the aquifers, especially the Lower Tertiary Sandy Aquifer, is a product of evapotranspiration prior to or during recharge and that the main source of the solutes is wet and dry precipitation. Although water rock interaction is important to solute content, it is not the reason the water becomes brackish. These conclusions are in keeping with the findings of another integrated groundwater study in the Tyrendarra Sub-basin (Bennetts, 2005).

7.4 Synthesis

Within this chapter original results of isotopic analyses of groundwater samples from the Otway Basin have been presented. The results include δ^2 H, δ^{18} O, δ^{13} C, 87 Sr/ 86 Sr, radiocarbon activity (%mC), 36 Cl/Cl and 36 Cl concentration data. These data provide evidence for the source of the groundwater and its solutes, the processes that occur during residence in the aquifers and the timing of recharge and length of residence. In most cases the findings from the isotope geochemistry also support and strengthen the conclusions of previous chapters.

First, the oxygen and hydrogen isotopic composition of the groundwater confirms that it is sourced from meteoric water. The exception to this is the one sample from the Upper Tertiary Carbonate Aquifer that has been affected by seawater incursion and is now confirmed to be a mixture of groundwater and seawater. The variation in the groundwater composition reflects modern seasonal variation and depleted and enriched samples are due this and also to geographic location, rather than past climate change. The samples located in the northern regions receive rainwater that is more enriched due to the latitude effect on precipitation. Other than the seawater intrusion noted above, the brackish groundwater samples have enriched and fractionated δ^2 H, δ^{18} O values that confirm evaporation is the dominant mechanism of salinisation.

The spread of the δ^2 H, δ^{18} O of the groundwater samples plots to the left of the LMWL and GMWL, indicating that fractionation has affected the water since recharge. This fractionation is the result of a combination of processes, namely interaction or dissolution of volcanogenic CO₂ and H₂S and silicate hydration through water-rock interaction. Carbonate mineral dissolution and precipitation in and out of groundwater has occurred in the past and is ongoing, including exchange of oxygen and carbon between the two phases, and has resulted in fractionation of the isotopes of these elements. The fractionation of the carbon is ongoing and is still evident in the δ^{13} C values of the water. However, the past interaction has resulted in formation of carbonate cements that have δ^{18} O values that are more or less in equilibrium with the groundwater δ^{18} O composition. Therefore, the oxygen of the groundwater is no longer strongly fractionated by the carbonate interaction and this process is not responsible for the resident groundwater plotting to the left of the LMWL.

In addition to interaction with carbonate minerals, the carbon isotope composition of the DIC of the groundwater reveals other carbon sources and processes. In some samples, especially close to recharge zones, the majority of the dissolved carbon is from dissolution of soil CO₂. The pH of the unsaturated zone, controlled to a large extent by the surficial geology, is a strong control on the δ^{13} C value of the DIC from soil CO₂. Subsequent to recharge, carbon is added to solution from carbonate minerals and interaction with volcanic CO₂, both of which enrich ¹³C in the DIC, making them hard to distinguish from one another. The addition of volcanic CO₂ to the system affects the CaCO₃-Ca⁺-HCO₃ equilibrium, possibly allowing the continued dissolution of carbonate minerals where the groundwater would otherwise be close to saturation with respect to calcite. Hence, the volcanic CO₂ may be an effective catalyst for karstification in the Upper Tertiary Carbonate Aquifer. The presence of natural gas currently escaping from depth or being trapped in caves near the surface has repercussions for the viability of CO₂ sequestration in the basin.

Carbonate mineral interaction is also a strong control on the ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ ratios measured from the groundwater, secondary only to the inherited ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ composition of precipitation. A trend of carbonate interaction is visible in the siliciclastic Lower Tertiary Sandy groundwater with increasing depth, supporting the conclusions regarding calcite equilibrium above. Dissolution of strontium from the rocks of the Pliocene-Quaternary Volcanic Aquifer also lowers the ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ ratio in several samples, including some from the Plio-Pleistocene Heterogeneous and Upper Tertiary Carbonate Aquifers. This finding confirms the interaquifer flow paths and recharge paths via eruption centres delineated in previous chapters. No evidence of interaction with silicate minerals other than those of the volcanic rocks is found within the ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ data. Therefore, the silicate hydration that has been identified as contributing to oxygen fractionation must occur mostly within the Pliocene-Quaternary Volcanic Aquifer.

Using radiocarbon activity in the groundwater, the age of the water was calculated. Based on the conclusions detailed above, it was corrected for addition of dead carbon from volcanic CO_2 and marine carbonate minerals in the aquifer matrices. The dilution correction factor utilised the δ^{13} C of the DIC and some generalised parameters for recharge conditions. The ¹⁴C ages confirm there is significant interaction between the Lower Tertiary Sandy Aquifer and the Upper Cretaceous Aquifer in the Gellibrand area, where younger water from the latter mixes with older water in the former and causes a decrease in the groundwater age down-gradient. Groundwater from the Lower Tertiary Sandy Aquifer at Dartmoor has some component of water containing bomb-pulse ¹⁴C, the addition of which is probably aided by rapid infiltration via karst features in the overlying Upper Tertiary Carbonate Aquifer. The seemingly anomalously high groundwater ages of the samples from the Clifton Formation are partly due to induced leakage of older water from adjacent aquitards as a result of pumping (Chapter 5). Lastly, there are several groundwater ages for the Lower Tertiary Sandy Aquifer that are greater than 22,000 years. These samples all have low salinity (EC $\approx 1,000-1,500 \ \mu\text{S/cm}$), confirming the theory that the recharge environment prior to the last glacial maximum (c. 20 ka) was a favourable one due to a higher P/E ratio.

Most of the conclusions made based on the radiocarbon data regarding the duration of groundwater residency and the timing of recharge are supported by the ³⁶Cl data. Application of the groundwater data along with expected atmospheric and hypogenic ³⁶Cl inputs to an equation for residence time resulted in negative values and confirmed that the groundwater is younger than 100,000 years. Some locations along flow paths show a decrease in ³⁶Cl/Cl and other paths show an increase, indicating that neither hypogenic addition nor decay of the radioactive isotope are the dominant process controlling its distribution. However, the relationship between groundwater ¹⁴C age and ³⁶Cl/Cl match well with the fluctuations in ³⁶Cl production in the atmosphere over time. Therefore, it is concluded that the change of R₀ in the past has caused the variable concentration of ³⁶Cl in the groundwater. The result suggests that the samples that were too old to be dated by radiocarbon (Port Fairy and Koroit) may be more than 37,000 years old and, therefore, pertain to the pluvial period between the last interglacial and the last glacial maximum.

With the ³⁶Cl/Cl ratios in the groundwater being close to those of rainfall, and generally increasing with age, the source of the dissolved chloride is confirmed to be cyclic rather than old connate salt. This is in line with the observation that the O and H isotopic composition of the groundwater shows no mixing with remnant connate water.

Chapter 8

Conclusions

This study provides a robust conceptual model of the regional physical and chemical hydrogeology of the Otway Basin, incorporating details, where relevant, of the region's geology, climate, groundwater and land use. The groundwater resources of the Otway Basin are both voluminous and valuable and this thesis fills a knowledge gap that has been noted by stake holders, hydrogeologists and managers alike. This concluding chapter provides an overview of the findings of this study by setting out the integrated picture of the Otway Basin region that is incrementally established throughout the preceding chapters. This discussion of the study's conclusions is grouped around four key themes which correspond to the objectives of the study (Chapter 1): geological controls on flow; recharge, discharge and storage characteristics; palaeoclimate influence on groundwater; and hydrogeochemical evolution.

8.1 Geological controls on flow

In this thesis, the sedimentary and volcanic rocks of the basin are organised into regional hydrostratigraphic units based on current knowledge of their lithology, porosity and hydraulic conductivity. These units are mapped in a 3D model that can be utilised in future numerical modelling endeavours. Isopach and structure contour maps are derived from the 3D model and provide new insight into the subsurface morphology of the hydrostratigraphic units. In turn, it is demonstrated that this morphology is an important influence on the groundwater flow.

For example, the differing geology of the sub-basins affects the groundwater velocity and residence times. Radiocarbon ages indicate that the Tyrendarra Sub-basin hosts the oldest water of the Lower Tertiary Sandy Aquifer. According to ³⁶Cl values, the water in the east of this sub-basin may have been recharged around 37,000 years ago. This central zone receives its water from flow through thick overlying sequences of basalt, limestone and marl. In contrast, the confining marl aquitard is significantly thinner in the Gambier Sub-basin to the west and is absent in the far east of the basin (parts of the Port Campbell and Barwon Downs Sub-basins; Figure 2.14). These eastern and western areas also lack basalt cover and therefore allow for more rapid infiltration to the Lower Tertiary Sandy Aquifer and have correspondingly lower ¹⁴C residence times.

A second example of geological influence on groundwater flow is the important effect of the lithofacies variability within the Pliocene-Quaternary Volcanic Aquifer on recharge mechanisms and groundwater drainage divides. A detailed analysis of the groundwater flow patterns at different depths and scales is undertaken through the creation of a watertable map and 4 potentiometric maps from density-corrected hydraulic head data (Chapter 4). The analysis confirms that the later phase eruption points, which are mostly phreatic or phreato-magmatic, provide preferential pathways for infiltration to the watertable because of high hydraulic conductivity. It is suggested here that this characteristic could also apply to infiltration of deep aquifers via vertical flow through volcanic conduits, such as diatremes.

8.2 Recharge, discharge and storage characteristics

One of the aims of this study is to map the contemporary groundwater flow systems and the recharge and discharge zones of the main aquifers. The potentiometric maps largely fulfill this aim and are complimented by additional mapping of the groundwater flow along 3 cross sections (Chapter 5). This allows for the identification of the main features of the flow systems, and when integrated with other resources, such as meteorological and oceanographic data, provide a physical hydrogeology framework of the basin.

The new maps and sections presented in this thesis illustrate that the relationship between surface water bodies, shallow groundwater and the underlying aquifers and aquitards is through an interaction between local-, intermediate- and regional-scale flow systems. Local-scale flow systems are complex, with flow lines often being radially oriented around topographic features. They have shorter groundwater residence times compared to regional systems. Regional flow in the aquifers is from the inland basin margins towards the coastline. Regional recharge zones are generally inland and are both diffuse (either through overlying units or directly to the aquifer via exposed areas) and point source (e.g. karst features, volcanic vents or faults).

Regional discharge from the Upper Tertiary Carbonate Aquifer is close to the coast, both onshore and offshore. Seawater has intruded into this aquifer at Port Macdonnell, Port Fairy and Warrnambool and mixed with groundwater, which is confirmed at the first location by the enriched δ^2 H and δ^{18} O values of the groundwater. Analysis of these zones indicates that the freshwater-saltwater interface in this aquifer is in a transient state of upward/landward movement as a result of onshore hydraulic head loss. The head loss is due to the combined effect of extraction and recent climate change and constitutes a decline in aquifer storage. These same factors are also behind the recent drying of volcanic lakes and the non-steady state of all the aquifers. Watertable fluctuation analysis of hydrographs indicates that winter recharge does not exhibit a declining trend from season to season, but a steady decrease in net recharge is due to the increased loss of water to evapotranspiration. The trends of rainfall and temperature corroborate this finding. Future climate change is likely to exacerbate this situation and may in turn contribute to a decline in groundwater quality.

Regional discharge from the Lower Tertiary Sandy Aquifer is submarine, via seepage through the overlying units and by cyclical mixing with seawater within the aquifer at the interface. The freshwater-saltwater interface is situated at a yet-unknown position offshore. Low salinity water is identified at random levels and locations within the oceanic water column from electrical conductivity profiles and this water could represent submarine groundwater discharge. It is postulated here that the freshwater-saltwater interface in the Lower Tertiary Sandy Aquifer is also in a transient state of upward or landward movement. The movement is a lagging aquifer response to current potentiometric conditions, i.e. interface migration from a previous position in equilibrium with the low sea level during the last glacial maximum is inferred. However, the movement may be accelerated by development of cones of depression at the coastal extraction sites of Portland and Port Fairy and the irrigation area between Robe and Beachport.

8.3 Palaeoclimate influence on groundwater

Chapter 3 focuses on the current climate and palaeoclimate of the Otway Basin region. A detailed analysis of the climatology, both past and present, is warranted in response to the project aim of placing the contemporary groundwater flow system within the context of the palaeohydrology regime. Although primary climatology data is not collected as part of this study, the major contribution of Chapter 3 is the integration of numerous forms of climate evidence to form a conceptual model of the recent evolution of the entire region. When married with the understanding of the flow system gained through other methods (Section 8.2 above), this model further illuminates certain aspects of the groundwater flow system and its evolution.

In short, the findings clearly show that changes in the precipitation/evaporation ratio (P/E ratio) over the last 100,000 years have affected the hydrological balance of the region. These changes, combined with eustatic fluctuations and the dramatic shifts in surface drainage patterns due to the eruption of voluminous lava flows, have contributed to the current physical and chemical state of the groundwater.

This study identifies the concentration of solutes by evapotranspiration near the surface as the most important factor affecting groundwater salinity (Section 8.4 below). Therefore, fluctuations of the P/E ratio over time are important in controlling the groundwater quality. By tracking the palaeolimnology of the region and other climate proxies, the history of the P/E ratio is reconstructed. Specifically, lakes had begun to develop in the Otway Basin region before the early Pleistocene in response to the blocking of drainage lines by the basaltic deposits of the Pliocene-Quaternary Volcanic Aquifer. There is no conclusive evidence of the salinity of lake environments before the last interglacial at about 130 ka. However, after the aridity of the last interglacial, a pluvial period resulted in the expansion of existing lakes and development of new lakes, most of which were freshwater. Accordingly, this was a period of effective recharge of low salinity groundwater. Around the time of the last glacial (20 ka), and for some time afterwards, the climate became drier, lakes receded and saline environments developed at varying rates and degrees throughout the region according to the nature of the surface water drainage (e.g. discharge lake vs. through-flow lake). Consequently, groundwater recharge quality has been affected by this change.

The best example of this connection between palaeoclimate and groundwater quality is seen in the Lower Tertiary Aquifer, although similar processes are also evident in the salinity distribution of the other aquifers. In some of the up-gradient areas of the Lower Tertiary Sandy Aquifer the groundwater is brackish; whereas the corresponding downgradient areas host older, fresher groundwater. This is an indication that the recharge water has slowly increased in salinity over time. Carbon-14 dating of the water in this aquifer supports the age relationship between the up-gradient and down-gradient groundwater. Indeed, it provides new insight into the spatial distribution and timing of development of surface water and shallow groundwater salinity. The region of the Lower Tertiary Aquifer that seems immune to these climatological changes is in the far east of the basin, near the Otway Ranges. The lower temperature and higher rainfall of this mountainous region has ensured that evaporative concentration has not affected the infiltrating groundwater to the same degree.

8.4 Hydrogeochemical evolution

During the course of this study in excess of 100 boreholes were sampled for geochemical analyses on the groundwater. Some analytical methods are applied to the water of the basin for the first time, for example, no groundwater from the Otway Basin had been analysed for its chlorine-36 signature prior to this study. As such, these new data, particularly the isotope geochemistry data, are an important contribution to the current knowledge of the hydrogeology of southeast Australia. The integration of the results of the various analytes with each other, and with the developed understanding of the flow system, produces several conclusions regarding the hydrogeochemical evolution of the groundwater of the Otway Basin.

There are two ways in which the Recent volcanic rocks have had an effect on the chemical composition of the groundwater: through water-rock interaction and water-gas interaction. These processes are confirmed by corroborating evidence from major ion, minor ion, and stable isotope geochemistry. Because the Pliocene-Quaternary Volcanic Aquifer covers much of the surface area of the Otway Basin, water infiltrating any of the other aquifers has often already interacted with basalt or tuff. This interaction has resulted in a complex suite of silicate mineral weathering reactions and the subsequent formation of clay minerals and occurrence of cation exchange. Degassing from magmatic sources has allowed interaction with, or dissolution of CO_2 , H_2S , HF, B and possibly other elements. The O and H isotopes of the water have been fractionated by this process and consequently, the groundwater composition lies to the left of the local meteoric water line. These data also reveal that the groundwater currently hosted in the aquifers has not interacted with magmatic water or H_2O_{qas} .

However, not all areas of the Otway Basin are blanketed by basalt. The groundwater of the Gambier Sub-basin is much more strongly influenced by the carbonate lithology of the Plio-Pleistocene Heterogeneous Aquifer. Accordingly, the Lower Tertiary Sandy Aquifer can exhibit a $Ca^{2+}HCO_3^-$ type water in some parts of this region. Dissolution of carbonates occurs in all the aquifers to varying degrees, even in siliciclastic aquifers. This is confirmed by the decrease of ${}^{87}Sr/{}^{86}Sr$ values in the groundwater below the rainfall signature. It has been a long-lived process in some aquifers, as evidenced by significant and widespread karst features. The ongoing input of CO_2 below the soil zone from volcanic sources allows continued dissolution of carbonate of the aquifer matrix and other weathering reactions where they otherwise might be impossible due to saturation indices exceeding zero.

Some mature water of low salinity in the Lower Tertiary Sandy Aquifer develops its characteristic $Na^+HCO_3^-$ type because of this input of volcanic CO_2 . The bicarbonate is derived from consumption of CO_2 in various weathering reactions, involving both carbonates and silicates. Even though all major cations are added to solution by such water-rock reactions, sodium tends to dominate in this case, due to cation exchange favouring its release to solution at such salinity levels.

In contrast, high salinity or brackish water in most of the aquifers, including the Lower Tertiary Aquifer, is Na⁺Cl⁻ type. This water is in flow systems where it has experienced both long and short residence times, which is one indication that water-rock interaction is not the main mechanism of accumulation of the solutes. In the case of long residence times, it has been previously suggested that the water or its solutes were derived from connate marine water, also Na^+Cl^- type. This is now shown to be incorrect by use of oxygen, hydrogen and chloride isotopes. This study confirms what others have also proposed, namely that the process of salinisation is dominantly evapotranspiration, concentrating the cyclic salt which is mainly sodium chloride.

8.5 Future of the resource

In summary, this study builds a conceptual model of the physical and chemical hydrogeology of the Otway Basin which can be used to inform future work in securing the sustainability of the groundwater resources of the region. Because of this comprehensive understanding of the system, these resources are now known to be vulnerable to certain future risks. Specifically, the depletion and salinisation driven by future climate change is an important risk and will require careful management of recharge zones. Additionally, although the timing is not constrained, the freshwater-saltwater interface can be expected to migrate landward and/or upward in the unconfined and confined aquifers and therefore contamination of valuable coastal groundwater resources may occur. Continued head loss in aquifers due to groundwater extraction is also anticipated and the resulting hydraulic gradients may induce different interaquifer flow regimes in the future. This study provides the base scientific understanding, both physical and chemical, of the groundwater flow systems; a model against which findings from further inquiries may be compared.

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