ORIGINAL ARTICLE

Use of a conceptual hydrogeological model and a time variant water budget analysis to determine controls on salinity in Lake Burrumbeet in southeast Australia

Yohannes Yihdego · John A. Webb

Abstract Lake Burrumbeet, which is located in the basalt plains of western Victoria, Australia, has a much lower salinity than other lakes in the region. To determine the factors causing the freshness of this water body, an assessment was carried out using existing geological information to reconstruct the palaeotopography beneath basalt flows, and using hydrogeological data, bore hydrographs and hydrological data together with a time variant lake water balance analysis calibrated from 1998 to 2008. The calibrated water balance of Lake Burrumbeet shows that groundwater outflow has a major influence on lake levels (31 % of water losses). The lake water budget analysis confirms that the interaction of Lake Burrumbeet with the surrounding groundwater system is volumetrically significant. Geological re-construction of the course of the buried river valley sediments (Deep Lead underlying the basalt) shows that they run directly beneath Lake Burrumbeet. Groundwater leakage into these sediments is sufficient to transfer enough salt from the lake to maintain a low to moderate lake salinity. The conceptual hydrogeological model of Lake Burrumbeet, coupled with the time variant water budget analysis, show that the difference in salinity among the lakes in the region can be explained by variations in the configuration and hydrogeological setting of the lakes, and this hypothesis provides new understanding of the relative freshness of the lake.

Introduction

In southwestern Victoria, a large number of lakes are scattered across the volcanic plains and many have problems with increasing salinity. Most of the lakes and wetlands are in small, shallow closed drainage basins which only receive limited runoff from the immediate surroundings, and are therefore dependent on rainfall and groundwater inflow. Groundwater is considered as an important component sustaining the water levels during summer (Tweed et al. 2009; Barton et al. 2013). Only a limited number of lakes in the study area (e.g. Lake Burrumbeet) receive substantial inflow from creeks. Many of the lakes in the region, including Lake Burrumbeet, were dry for the period 2005-2010, due to the prolonged period of below average rainfall from 1997 to 2010 as well as human activity such as the extraction of groundwater (Adler and Lawrence 2004; Dahlhaus et al. 2010; Yihdego and Webb 2012).

Shallow lakes and wetlands are almost invariably connected hydraulically to the surrounding unconfined aquifer systems (Winter 1981), and quantification of groundwater flux to/from a lake is generally necessary for an accurate lake hydrologic, chemical, nutrient and/or energy budget (e.g. Winter 1978; Kirillin et al. 2013; Sener and Davraz 2013; Yihdego and Becht 2013). However, the groundwater flux to a surface water body is one of the more difficult components of the water balance to measure (e.g. Harvey et al. 2000; Harvey and Sibray 2001; Demlie et al. 2007), particularly in geologically complex areas (Nelson and Mayo 2014; Ran et al. 2014). Darcian calculations of

Y. Yihdego · J. A. Webb
Environmental Geoscience, La Trobe University, Melbourne, VIC 3086, Australia

Y. Yihdego (✉)
Snowy Mountains Engineering Corporation (SMEC), Sydney, NSW 2060, Australia
e-mail: yohannesyihdego@gmail.com

groundwater discharge to surface water bodies are often limited by the very low hydraulic gradients in aquifers immediately adjacent to wetlands.

Groundwater–lake interactions are determined by the hydrogeological setting of the lake. Previous studies on the basalt plains of western Victoria (e.g. Barton et al. 2007, 2013; Jones et al. 2001; Yihdego and Webb 2012) have identified several different hydrogeological controls on the lakes there. Groundwater discharge (springs) at the break of slope has been identified as an important input to some lakes. Springs are typically located at the edge of younger, more permeable third phase basalts where these overlie older basalt phases with poor drainage capacities, and often feed lakes (e.g. Lake Terrinallum). Also, numerous lakes have formed along the edge of basalt flows in topographic depressions resulting from the diversion of creeks by the lava flows. In addition, the regional discharge of groundwater where shallow basement ridges form hydraulic barriers to lateral groundwater flow, e.g. along the southern margin of the buried river valleys of the Deep Lead system, has been identified as a major contributor to groundwater discharge and the formation of lakes in the central part of the study area (Raiber et al. 2009).

Coram (1996) studied the lake water and salt budget of lakes in western Victoria to identify the interaction between the lakes and the groundwater system, and identified some lakes as through-flow lakes (Lakes Murdeduke, Bookar, Colongulac and Gnarpurt) and others as discharge lakes (Lake Beeac). The differences are due to variations in the local hydrostratigraphy downgradient of the lakes, particularly differences in the relative proportion of groundwater outflow. The lakes in this area show a range of salinities despite being located in an area with similar climatic conditions, topography and distances to the ocean. It is speculated that geological setting could explain the variation in salinity among these lakes (Yihdego and Webb 2012).

In this study, Lake Burrumbeet was selected for a water budget analysis because it is relatively fresh and has good records of lake levels (1998–2007) and salinity (1991–2007), as well as gauged inflow (1975–2007) data. There has been no detailed previous work on the hydrology of Lake Burrumbeet, but there are a number of related and relevant studies in the region (e.g. Yihdego and Webb 2012, 2013). Dahlhaus et al. (2008) compiled historical records in relation to the water levels and water quality of Lake Burrumbeet. Learmonth (1853, in Bride 1898), the first European settler to visit Lake Burrumbeet, recorded the water as brackish in August 1837. When the lake was next visited in January 1838, it contained a few inches of intensely salty water and was unrecognisable from the previous visit. Withers (1870) noted that the following year (1839) the lake was dry and remained so for several

successive summers. In 1849 there was fresh water to the depth of at least 1.2 m (Dahlhaus et al. 2008).

This study aims to identify the controls on the salinity of Lake Burrumbeet, by integrating several lines of evidence, including a lake water balance analysis and lake-groundwater interactions, coupled with the reconstruction of the pre-basaltic land surface, using geological, geophysical, hydrogeological and related data. This analysis will provide a better understanding of the lake water level changes and associated salinity variations, and provide enhanced insight into the intricacies of the flow system around the lake.

Site description

Lake Burrumbeet is located in central western Victoria and is the largest of four shallow lakes in the region (Fig. 1), with an area of ~ 23 km^2, and is utilised for recreational boating, fishing and camping. Lake Burrumbeet lies within the upper Hopkins River catchment, which in this area is generally gently undulating with an average elevation of only 385 m Australian Height Datum (AHD), i.e. above sea level. Topographic lows are often associated with swamps or lakes. The lake catchment, which has an area of 298 km^2, is bounded in the north by the Great Dividing Range, which separates coastal streams from those flowing inland. The lake catchment area coincides with the Glenelg-Hopkins catchment boundary shown in Fig. 1.

The lake is fed by Burrumbeet Creek, the major inflow, from the east and Canico Creek from the south (Fig. 1). The lake outlet is situated on the southwest shore of the lake, and flows into Baille Creek, which is a tributary of Mt Emu Creek. The Lake Burrumbeet outlet structure is a 40 m wide weir consisting of a concrete slab with an outlet height of 378.7 m AHD (Lawson and Treloar Pty Ltd. 2003). The weir can be adjusted with a removable wooden plank to elevate the outflow level to 379.1 m AHD.

Hydrometeorological data are available for Burrumbeet Creek, the main inflow to Lake Burrumbeet, as part of a flood frequency analysis (Lawson and Treloar Pty Ltd. 2003).

In general the discharge of Burrumbeet Creek decreased from 1996 to 2007, due to the drop in rainfall over this time, as shown by the cumulative deviation from mean rainfall. The average flow in 2006/2007 ($\sim 6.5 \times 10^3$ m^3/day) was ~ 19 % of the long-term flow (35×10^3 m^3/day), which is less than flow during the previous drought (1982/1983; 7×10^3 m^3/day). There is a strong correlation between rainfall and discharge.

The catchment of Lake Burrumbeet supports wool and prime lamb production, beef cattle and cereal crops and oilseeds. Other land uses include intensive animal
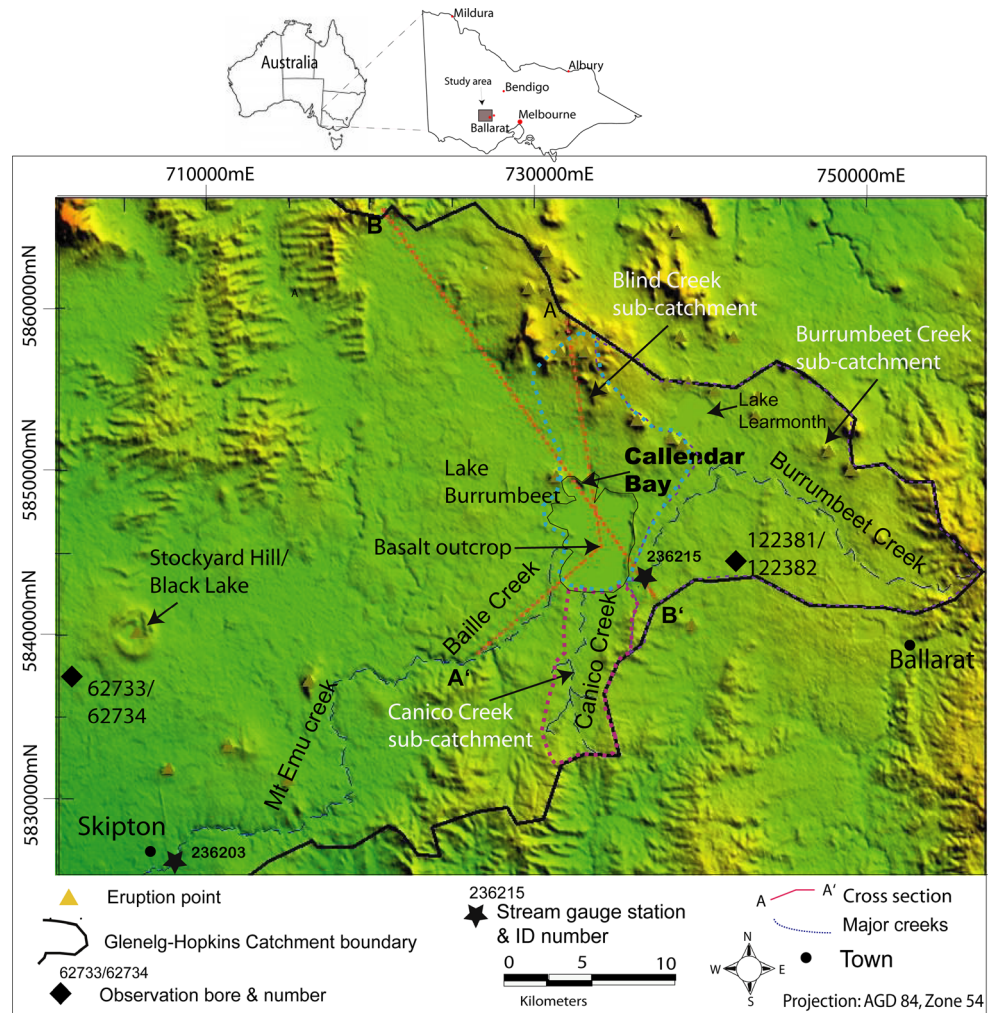
industries (pigs, horses) and blue gum plantations (Lawson and Treloar Pty Ltd. 2003).

Geological setting

Lake Burrumbeet catchment has a complex geology (Fig. 2). The Palaeozoic basement rocks (Cambrian and Ordovician sandstone and shale intruded by Devonian granites) outcrop over limited areas, and are overlain by an extensive cover of Cenozoic basalt, colluvium and alluvium. The airborne gravity image shows the subsurface extent of the buried granite intrusion beneath Lake Burrumbeet (Fig. 3).

Most of the catchment is covered by basalt and scoria of the Pleistocene Newer Volcanics, which erupted during three phases, ranging from about 6 million to a few thousand years ago. The second phase basalts, which erupted ~2–3 million years ago (Bennetts et al. 2003; Joyce 2003; Joyce et al. 2003), surround Lake Burrumbeet and have a thickness of 50–70 m, on average. Different flows are clearly evident, based on bore logs, separated by

weathering surfaces; the oldest flow appears to have filled a depression. The surface of the lavas around Lake Burrumbeet is deeply weathered with a 4–5 m thick regolith cover. The disruption of the drainage system by the basalt flows formed lakes and swamps, including Lake Burrumbeet.

Callendar Bay, a small semi-circular bay on the northwestern edge of Lake Burrumbeet (Fig. 2), occupies the site of an explosive eruption point (maar) that produced tuff and basaltic agglomerate, forming the nearby scoria cones, Mt Callendar to the west and a hill to the north east (Rosengren 1994). There are excellent exposures of these tuffs, showing cross-bedding and bomb sags, in the cliffs along the edge of the lake adjacent to Mt Callendar. Material weathered from the tuff forms beach gravels nearby, including blocks of granite, which were brought to surface as bombs during the volcanic eruption. Overlying the pyroclastics is a lava flow, also exposed in the cliffs around Callendar Bay.
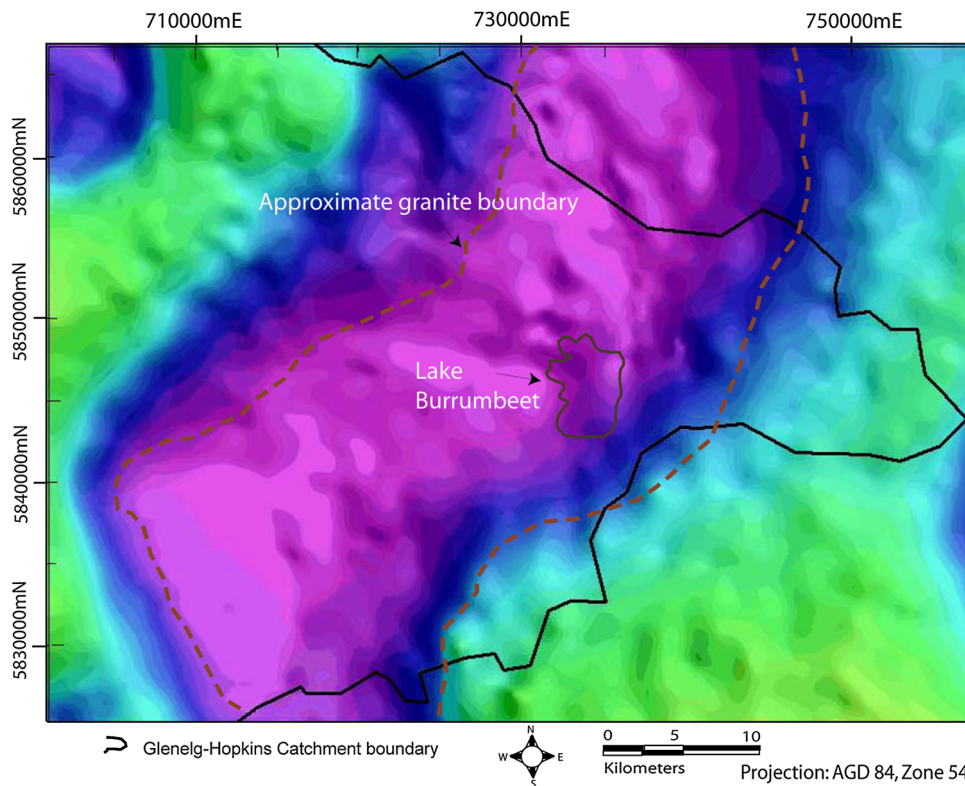
Underlying the basalts are the Early to Middle Miocene and Pliocene sands, ligneous clays and gravels of the

Fig. 2 Geological setting of Lake Burrumbeet area (after Geological Survey of Victoria)
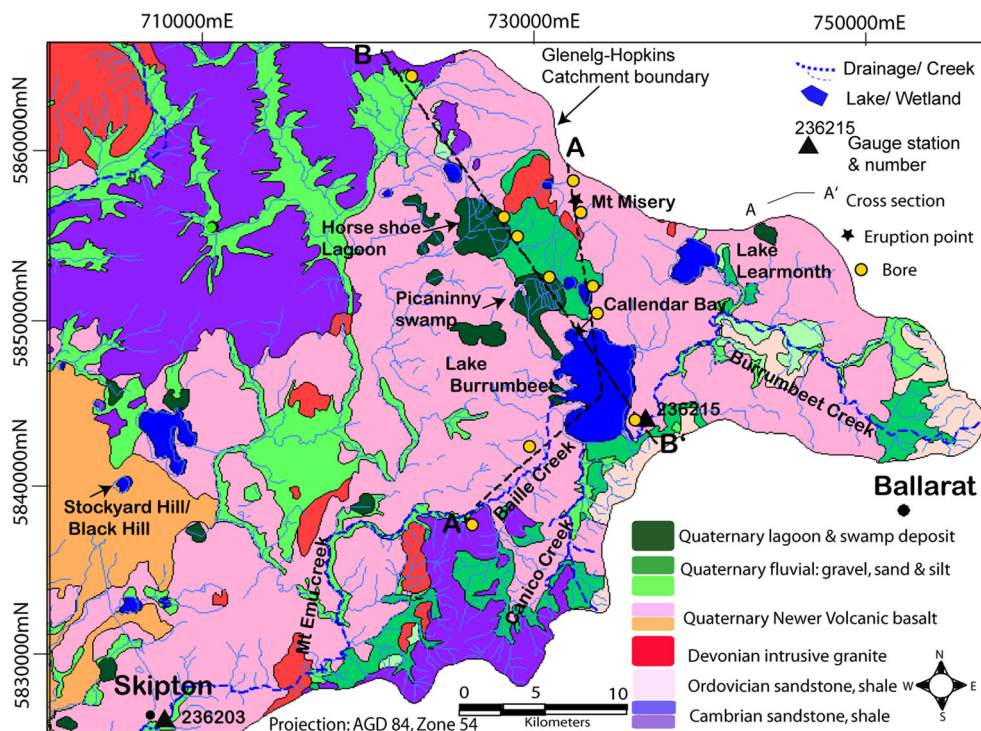
Fig. 3 Airborne gravity intensity image of Lake Burrumbeet area (*red* = high, *blue* = low) (courtesy Geological Survey of Victoria)

buried river valley sediments of the Deep Lead System (Calivil Formation), representing the pre-existing stream system incised deeply into a highly weathered early Cainozoic palaeo-surface (Holdgate et al. 2006; Raiber and Webb 2008). Deep Leads occur throughout Victoria; they are often important aquifers and are famous for their placer gold deposits (Raiber and Webb 2008).

The Deep Lead palaeodrainage system has a different distribution and flow direction to the present drainage on the basalt, because the drainage divide was shifted tens of

kilometres northward to its present position when the basalt eruptions built a new topography to a greater height than along the pre-existing drainage divide (Taylor and Gentle 2002).

The precise course of the Deep Lead system beneath the basalt has been difficult to determine (Taylor and Gentle 2002). In this study the reconstruction of the Deep Lead used detailed data from boreholes, mine records and regional mapping, as well as the basement palaeotopography, Deep Lead sediment thickness, basement outcrop and basalt thickness, to show that it lies to the south and west of the high-standing granite massif of Mount Misery (Fig. 4). The maximum thicknesses of Deep Lead sediment (Fig. 4) (up to 76 m) north and northeast of Lake Burrumbeet indicate the likely north-flowing outlets.

Hydrogeological setting

The relevant groundwater flow systems in the Lake Burrumbeet area (Dahlhaus et al. 2002) can be summarised as follows:

1. Local flow system in Quaternary alluvium (Fig. 4).
2. Intermediate and local groundwater flow system in fractured Palaeozoic basement (Fig. 4).
3. Regional and intermediate groundwater flow systems in the Newer Volcanics volcanic plain surrounding Lake Burrumbeet and underlying Deep Lead sediments (Fig. 4). There may also be local groundwater flow systems in the basalts around high points in the landscape such as scoria cones, with different flow directions to the regional flow path.

A palaeosoil horizon within the basalt succession probably acts as a confining layer between the basalt and the Deep Lead system (Fig. 4). It is most likely breached by the volcanic neck beneath the Callendar Bay maar, allowing hydraulic connection of Lake Burrumbeet with the underlying Deep Lead aquifer (Fig. 4). The volcanic neck represents a distinct hydro-stratigraphic unit within the Newer Volcanics lava flows, and probably has a higher hydraulic conductivity due to extensive fracturing. This unit may therefore comprise an important source of local recharge to both the shallow and deep groundwater systems of the lake area.

The volcanic plain and Deep Lead Systems have the most influence on the hydrology of Lake Burrumbeet hydrology. Little hydrogeological data exists for the buried portions of the basement, as few bores penetrate it, and no bore screens this aquifer in the lake area.

The potentiometric surface in the basalt aquifer was constructed using available bore data. The potentiometric surface decreases in elevation to the south west (Fig. 5) and is a subdued reflection of the topography, with high groundwater heads in the topographically elevated areas
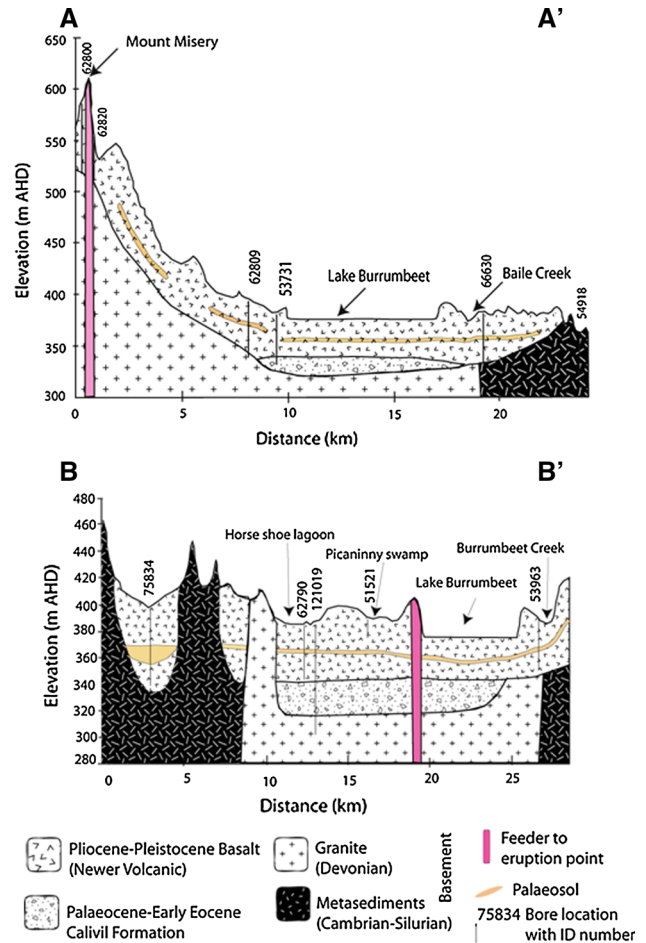

Fig. 4 Geological cross-sections *A-A'* and *B-B'*. Refer to Fig. 2 for cross-sections location

near the catchment boundary, and low groundwater heads in the valley floor along Burrumbeet Creek. Groundwater flow is from catchment boundaries towards Lake Burrumbeet and then southwest to the outlet, analogous to the surface drainage system (Fig. 1). Lake Burrumbeet intercepts part of this shallow ground water flow, which continues through the lake to the southwest, towards Baille Creek (Fig. 2). Thus Lake Burrumbeet is, to a large extent, an extension of the regional groundwater system.

The water table in the basalt shows significant seasonal fluctuations, e.g. 1.3–1.4 m in B122382 (Fig. 8), and these have been used to estimate recharge to the basalt aquifer.

Groundwater observation bores for the Deep Lead aquifer are generally too far apart to provide more than a regional framework of groundwater flow paths. Nevertheless it can be seen that the potentiometric surface of the Deep Lead slopes gently north with a hydraulic gradient of $\sim 2.6 \times 10^{-3}$ (Fig. 6), so the general flow direction is to the north, as expected from the pre-basaltic topographic gradient (Fig. 7), but opposite to both the present topography of the area and the potentiometry of the overlying
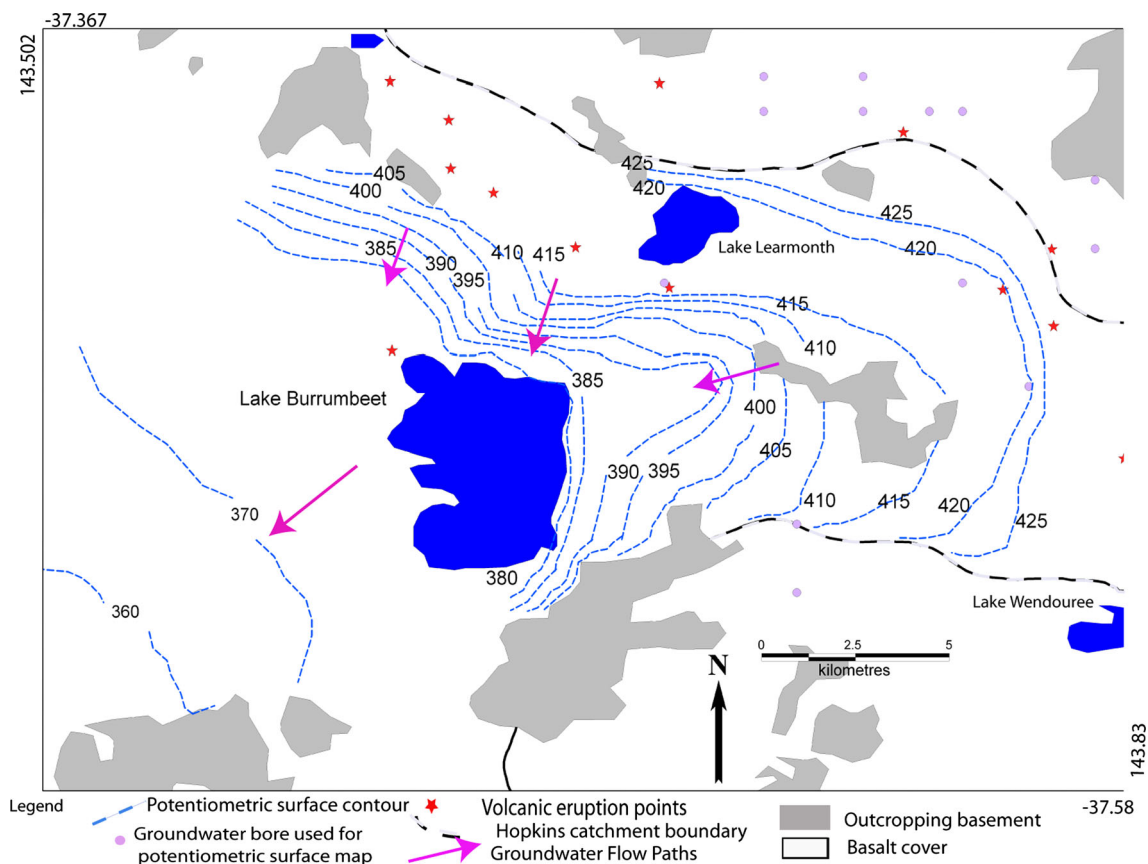
Fig. 5 Basalt aquifer potentiometric surface elevation

basalt water table aquifer (Fig. 5). Overall, the basalt and Deep Lead may be considered as separate aquifer systems with different hydraulic characteristics and recharge mechanisms (Yihdego and Webb 2011).

Comparison of potentiometric surface elevations in the basalt and Deep Lead at nested bore sites (Fig. 8) show that although both aquifers respond to recharge events similarly, the basalt hydrograph often has larger fluctuations (because it is an unconfined aquifer), which are relatively delayed in the Deep Lead aquifer. This reflects poor interconnection between the aquifers due to the thick palaeosoil separating the basalt flows and Deep Lead aquifer. The hydrographs (Fig. 8) indicate that there is a downward hydraulic gradient, with the potential for the Deep Lead aquifer to be recharged by basalt groundwater where the palaeosoil is breached in the vicinity of eruption points by the volcanic rocks beneath the volcanoes.

Water balance modelling

A water and salt balance model for Lake Burrumbeet was developed as a spreadsheet. The present study employs a modified difference water budget method, that improves the groundwater flux estimate using only Darcy's law by

linking the lake to the groundwater and incorporating the seasonal variation of the groundwater fluxes (Yihdego and Webb 2012). This method estimates the net groundwater flux by using the difference between the level of the lake and groundwater, along with the specific yield of the aquifer (from pumping test results from wells surrounding the lakes) and the area of the aquifer (obtained from hydrogeological maps). The reader is referred to Yihdego and Webb (2012) for the detail of the model.

The spreadsheet model used in this study has the capability to simultaneously solve the conservative solute mass balance along with the water mass balance of the lake, providing additional constraints on the water balance.

The initial Darcy's Law estimations were supplemented using a bucket model (described below) to capture seasonal net groundwater flux. The groundwater inflow and outflow estimates were then optimised during calibration of the model using the calculated net groundwater flux.
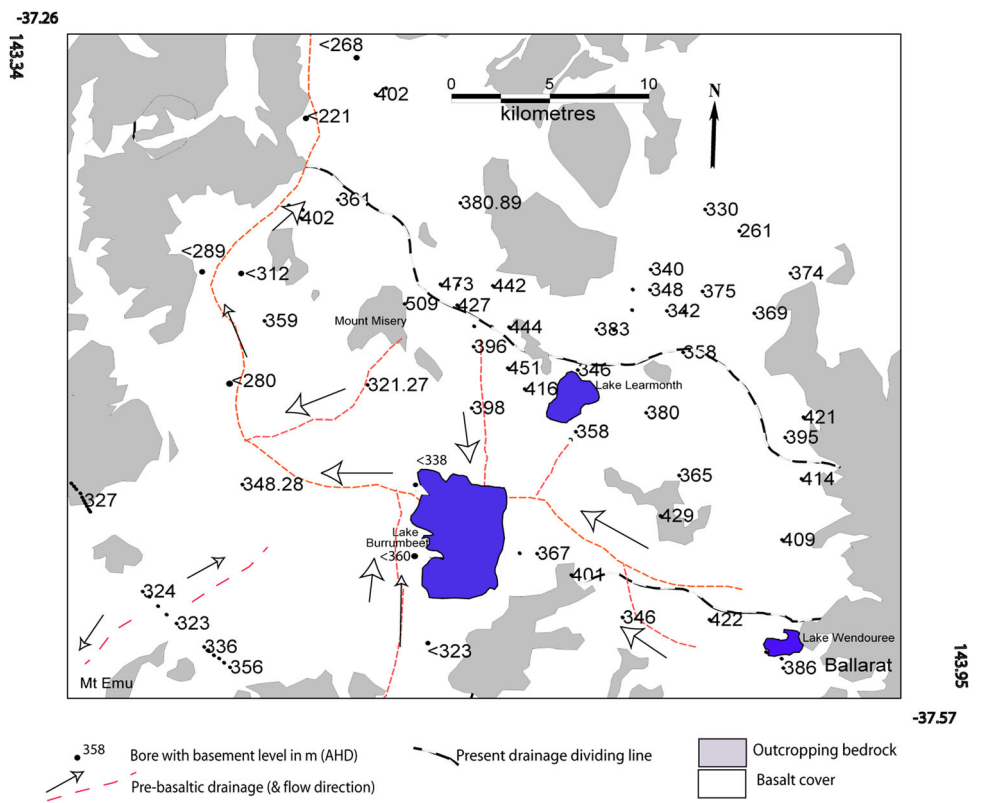
Groundwater fluxes

Darcy method Initially, for lake modelling, groundwater inflow/outflow from/to the lake was quantified using Darcy's Law (Eq. 1):
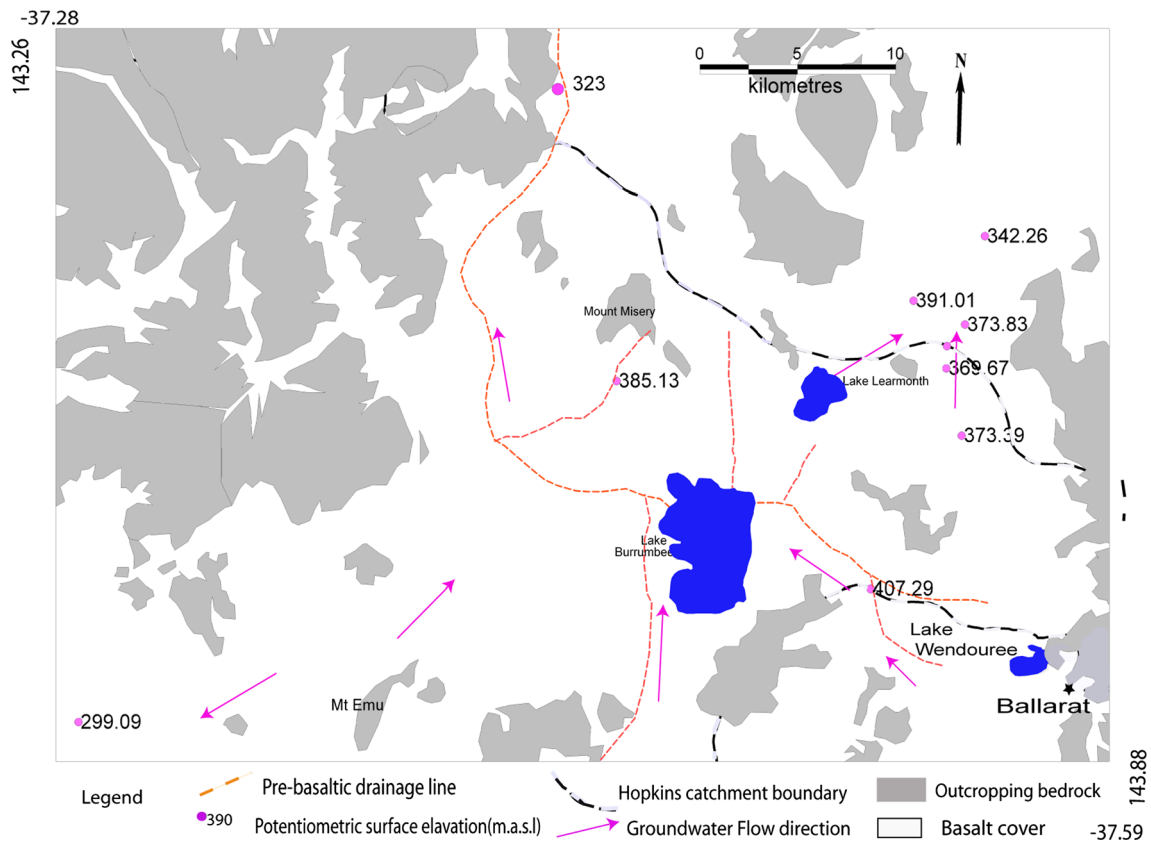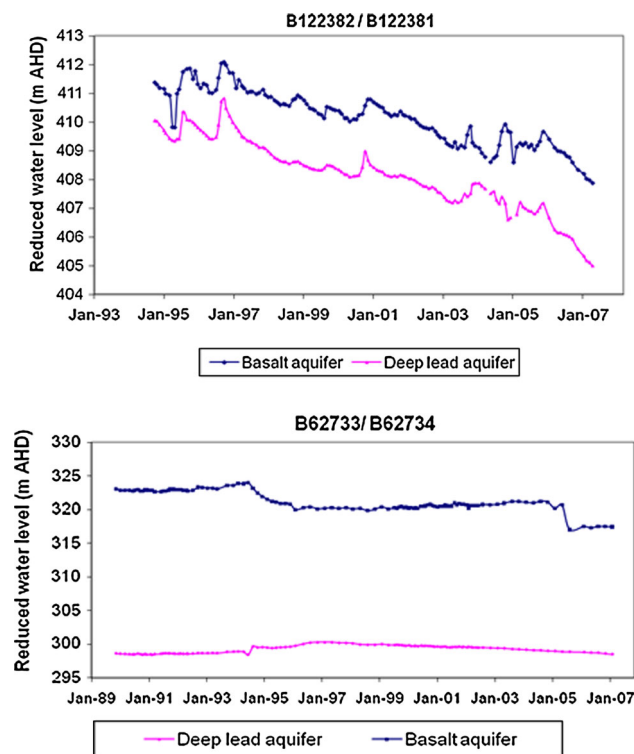
Fig. 6 Deep Lead aquifer potentiometric surface elevation

Fig. 7 Reconstruction of the sub-basaltic drainage system of the Lake Burrumbeet area, showing the basement palaeotopography

Fig. 8 Reduced water level in the basalt and Deep Lead aquifers at nested bore sites B122381/B122382 (5 km east of Lake Burrumbeet) and B62733/B62734 (29 km west of Lake Burrumbeet). Bore depths of 62733 and 62734 are 84 and 127 m, respectively. Bore depths of 122382 and 122381 are 45 and 104 m, respectively. The locations are shown on Fig. 1

$$G_{i,o(Darcy)} = K \times I \times A \tag{1}$$

where, G_i, $_o$ = groundwater inflow ($G_{i,\,Darcy}$) and outflow ($G_{o,\,Darcy}$) (m^3/month), K is the hydraulic conductivity of the aquifer/lake bed sediments/lake shore sediments (m/month), I is the hydraulic gradient between the aquifer and the lake, A is the cross-section area of the aquifer that is hydraulically connected to the lake (m^2).

Groundwater discharge/recharge to/from a lake occurs both through the lake bed and along the shore lines around the lake, so both were taken into account. The hydraulic conductivity values used were the lowest for the groundwater path being considered, i.e. lake bed sediments for the flux through the lake floor, and surrounding aquifer for the flux through the lake shore, because the sandy shoreline sediments have higher permeability than the basalt aquifer around Lake Burrumbeet.

Bucket model In the bucket model used to estimate net groundwater flux to/from the lake, the groundwater is lumped into an aquifer reservoir which is linked to the lake through the lake floor and lake shore sediments as follows:

$$G_{net} = (K_{bed} \times I_b \times A_{bed}) + (K_{shore} \times I_s \times A_{shore}) \tag{2}$$

where K_{bed} is the hydraulic conductivity of the lake bed sediments and K_{shore} is the hydraulic conductivity of the lake shore sediments, A_{bed} is the area of the lake bed and A_{shore} is the area of the lake shore, I_b is the hydraulic gradient between the aquifer and lake bed, I_s is the hydraulic gradient between the aquifer and lake shore.

Most groundwater flux to/from a lake occurs through the lake shore (Winter 1981; LaBaugh et al. 1997), because the permeability of the lake shore sediments and/or exposed surrounding aquifer is so much greater than that of the lake floor sediments, even though the area of the lake floor is greater than that of the shore line. Eq. (2) can be rearranged as:

$$G_{net} = C \times (\Delta h) \tag{3}$$

where G_{net} = net groundwater flux between the aquifer and the lake (m^3/month), C = conductance of the lake bed and lake shore Sediments. In the Lake Burrumbeet modelling, K_{shore} was replaced by K_{basalt}, for the reason outlined above. Separating the aquifer and the lake (m^2/month), Δh = head difference between the lake surface and the potentiometric surface of the hydraulically connected aquifer directly beneath and either side of the lake (m).

The value of C is given by;

$$C = (K_{bed} \times (A_L)/M) + (K_{shore} \times (A_{shore})/N), \tag{4}$$

where M is thickness of the lake bed sediments and N is thickness of the lake shore sediments. In the Lake Burrumbeet modelling, K_{shore} was replaced by K_{basalt}, for the reason outlined above.

To determine the head difference (Δh) between the aquifer and the lake, the groundwater level in the aquifer surrounding the lake is not taken from bore hydrographs but calculated from the change in groundwater storage in the aquifer:

$$G_{net} = S_y \times A_{AQ} \times \Delta H_{AQ} \tag{5}$$

G_{net} = net groundwater flux between aquifer and lake (m^3/month), S_y = storage coefficient (specific yield) of the aquifer (obtained from pumping tests or estimates of effective porosity), A_{AQ} = area of aquifer hydraulically connected to the lake, i.e. the area of the aquifer in which the groundwater level fluctuates due to interaction with the lake (m^2), ΔH_{AQ} = change in the groundwater level within the aquifer due to groundwater flow to/from the lake (m).

Equation 5 can be rearranged to give the change in the level of the potentiometric surface in the aquifer for any month (ΔH_m):

$$\Delta H_m = G_{net,m}/(S_y \times A_{AQ}) \tag{6}$$

This can be used to calculate the hydraulic head in the aquifer in any month using that in the previous month:

$$H_{m+1} = H_m + \Delta H_m \tag{7}$$

To run the model, the initial head in the aquifer is assumed the same as the lake level, implying no hydraulic gradient.

Combination method For the water budget modelling in this study, the total groundwater inflow/outflow was estimated using a combination approach, in which the net seasonal groundwater flux estimated using the bucket model (Eq. 3), supplements the Darcy's Law approach.

If G_{net}, using the bucket model (Eq. 5) is positive, groundwater inflow exceeds outflow, i.e. there is a net seasonal groundwater flow from the aquifer into the lake. This value can be considered the minimum amount of groundwater inflow to the lake, and is added to the estimated groundwater inflow using Darcy's Law:

$$G_i = G_{i(Darcy)} + G_{net} \tag{8}$$

$$G_o = G_{o(Darcy)} \tag{9}$$

If G_{net} from the bucket model is negative, groundwater outflow exceeds inflow, i.e. there is a net groundwater flow from the lake to the aquifer. This can be considered the minimum amount of groundwater outflow in the lake water budget, and is added to the groundwater outflow estimated using Darcy's Law:

$$G_i = G_{i(Darcy)} \tag{10}$$

$$G_o = G_{o(Darcy)} + G_{net} \tag{11}$$

Change in lake storage

The change in lake storage ($\Delta S(t)$) was calculated on a monthly basis from the precipitation ($P(t)$), evaporation ($E(t)$), surface inflow ($S_i(t)$), surface outflow ($S_o(t)$), abstraction due to pumping ($Ab(t)$), and the adjusted groundwater inflow ($G_{i,}$) and outflow (G_o) during that month:

$$\Delta S(t) = P(t) - E(t) + S_i(t) + G_{i,} - G_o - Ab(t) - S_o(t-1) \tag{12}$$

The water budget terms are expressed in volumetric units (e.g. precipitation and evaporation multiplied by the average lake surface area during that time).

The lake volume ($S(t)$) is updated for every time step with the lake volume change:

$$S(t) = S(t-1) + \Delta S(t) \tag{13}$$

where $S(t-1)$ corresponds to the lake volume from the previous month:

The calculated lake volume for any month is then compared to the actual lake volume for that month, calculated from the measured water depth.

Water level and salinity data were measured from 1998 to 2008. Monthly rainfall and evaporation values were obtained from the nearest weather station. For Lake Burrumbeet there is a gauging station on the main stream feeding the lake, and inflow from minor streams was calculated using catchment area ratioing (Yihdego and Webb 2012).

For the modelling it was necessary to convert the measured lake depths to lake volumes. Bathymetry data is not available for the lake; however, the relationship between lake area and depth was established by measuring the area on 12 Landsat images taken between 1972 and 2004, plotting this against the measured lake depth in the month when the images were taken, and determining a line of best fit. There is a polynomial relationship between lake area and lake depth, due to the lake's relatively flat bed and steep banks.

Groundwater flux estimation

Groundwater flow through lake shore Potentiometric surface contours for the basalt aquifer (Fig. 5) indicate that Lake Burrumbeet is a groundwater through-flow lake, with groundwater flow mainly entering the lake from the north and east side ($\sim$9 km perimeter) and leaving from the southwest perimeter ($\sim$3 km).

For estimation of the groundwater inflow using Darcy's Law, the cross-sectional area was calculated by multiplying the width of the groundwater inflow zone (9 km) by its average depth (14 m), which is the saturated thickness of the basalt aquifer above the palaeosol (Fig. 4). The hydraulic gradient, derived from potentiometric contours (Fig. 5), is on average 0.0032. Permeability in the basalt varies from tight fractures to open fractures and lava tubes, so the hydraulic conductivity is very variable. A value of 0.5 m/day was used, i.e. the average for basalts in the very similar upper Loddon catchment to the north (Hagerty 2006); both areas have basalt flows of the same age separated by clay-rich weathering surfaces.

Groundwater outflow was estimated in the same way as groundwater inflow, using the width of the groundwater outflow zone (3 km) and the hydraulic gradient, derived from potentiometric contours on the downgradient side of the lake (Fig. 5), averaging 0.0026. The groundwater outflow through the lake shore is set to zero when the lake dries out.

Groundwater flow through lake bed There are no values available for the permeability of the lake floor, but it is covered with a layer >1 m thick of clay derived from basalt weathering. Dahlhaus et al. (2002) estimated that clays in basalt weathering profiles in western Victoria typically have K values from 8.64×10^{-7} to 8.64×10^{-2} m/day.
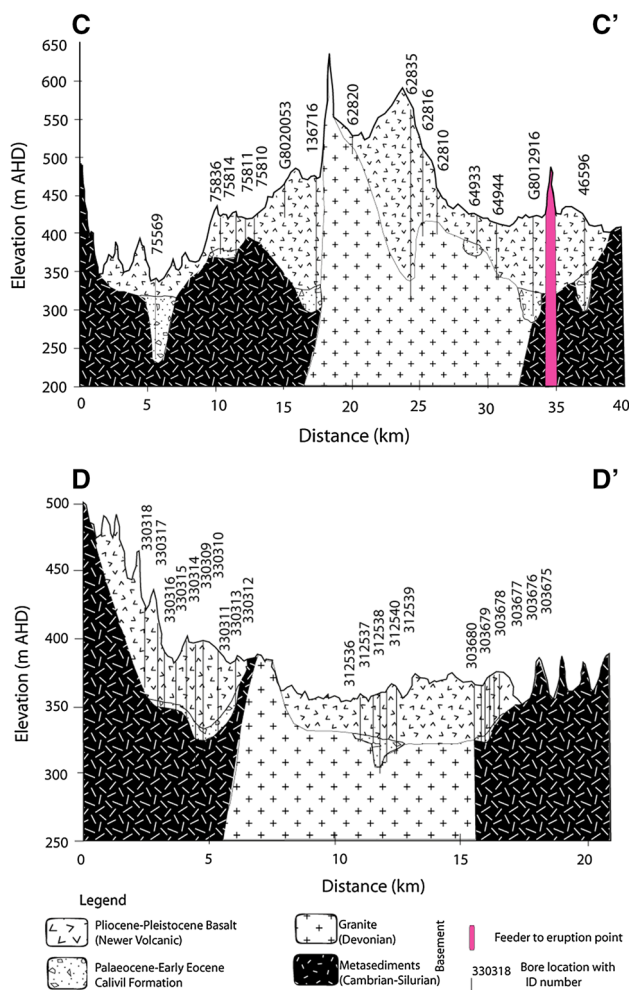
Fig. 9 Geological cross-sections. The positions of cross-section lines are shown on Fig. 10

Hence an average value of 8.64×10^{-4} m/day has been chosen for the permeability of the lake floor in order to estimate the groundwater outflow through the lake bed. The cross-sectional area (the lake floor covered by water) and the vertical hydraulic gradient (derived by comparing nearby bore hydrographs with the lake level) are estimated as ~ 6 km^2 and 0.0008, respectively. The groundwater outflow through the lake bed is set to zero when the lake dries out.

The monthly groundwater inflow and outflow volumes to/from Lake Burrumbeet through both the shore lines and lake floor, as discussed above, were estimated as $\sim 6 \times 10^3$ and $\sim 1.7 \times 10^3$ m^3, respectively. However, the groundwater outflow may be greater, because a volcanic neck (Callendar Bay) is found at the northwestern edge of the lake, and the lake floor is not entirely covered by clayey sediments; there are areas covered by basalt lava (Fig. 1). These areas may have values of hydraulic conductivity many orders of magnitude higher than the lake

bed clays. These were optimised during calibration of the model.

Model calibration

Model calibration has been carried out by adjusting the input parameters so that the simulated lake levels and salinities fit the observed lake levels and salinities. The initial model output showed simulated lake levels lying at higher levels than the observed lake levels, signifying a need for additional discharge.

The model was calibrated by minimising the sum of squared differences between observed and simulated lake levels. A "solver" optimization function in Excel was run and the parameters were iteratively adjusted in order to minimise the objective function, which is the sum of the squared differences between the observed and calculated lake levels/lake salinities (i.e. residual error). For the net groundwater flow estimation, initial specific yield and hydraulic conductivity values based on previous estimates and published values for similar material were adjusted during calibration. The optimised specific yield compares well with previous specific yield estimates of the basalt aquifer, ranging up to 0.18 (Dahlhaus et al. 2002; Bennetts et al. 2006; Raiber et al. 2009).

The best possible fit of the water budget model was attained with $\sim 37 \times 10^3$ m^3 and 355×10^3 m^3 monthly average groundwater inflow and outflow, with the exception of dry periods. These estimates suggest that there is substantial leakage from the lake, probably through the sand beaches in the caldera area down into the volcanic neck and underlying Deep Lead sediments (Fig. 4), away from the lake, following the course of the pre-basaltic drainage/ Deep Lead (Figs. 7, 9, 10). There may also be leakage through the basalt outcrops in the centre of the lake floor (Fig. 1).

Results and analysis

Lake water budget

The lake levels predicted by the water budget show good agreement with the measured lake level data, with an r^2 value of 0.96, which is acceptable within the range of errors associated with the water budget. The sum of the squared differences between observed and calculated lake levels does not exceed 0.12 m. Overall the model predicts the recession in lake level accurately (Fig. 11), with only minor discrepancies in dry periods due to overestimation of lake filling from surface inflows, because inflows into the lake at these times infiltrate directly into the dry lake-bed rather than filling the lake.
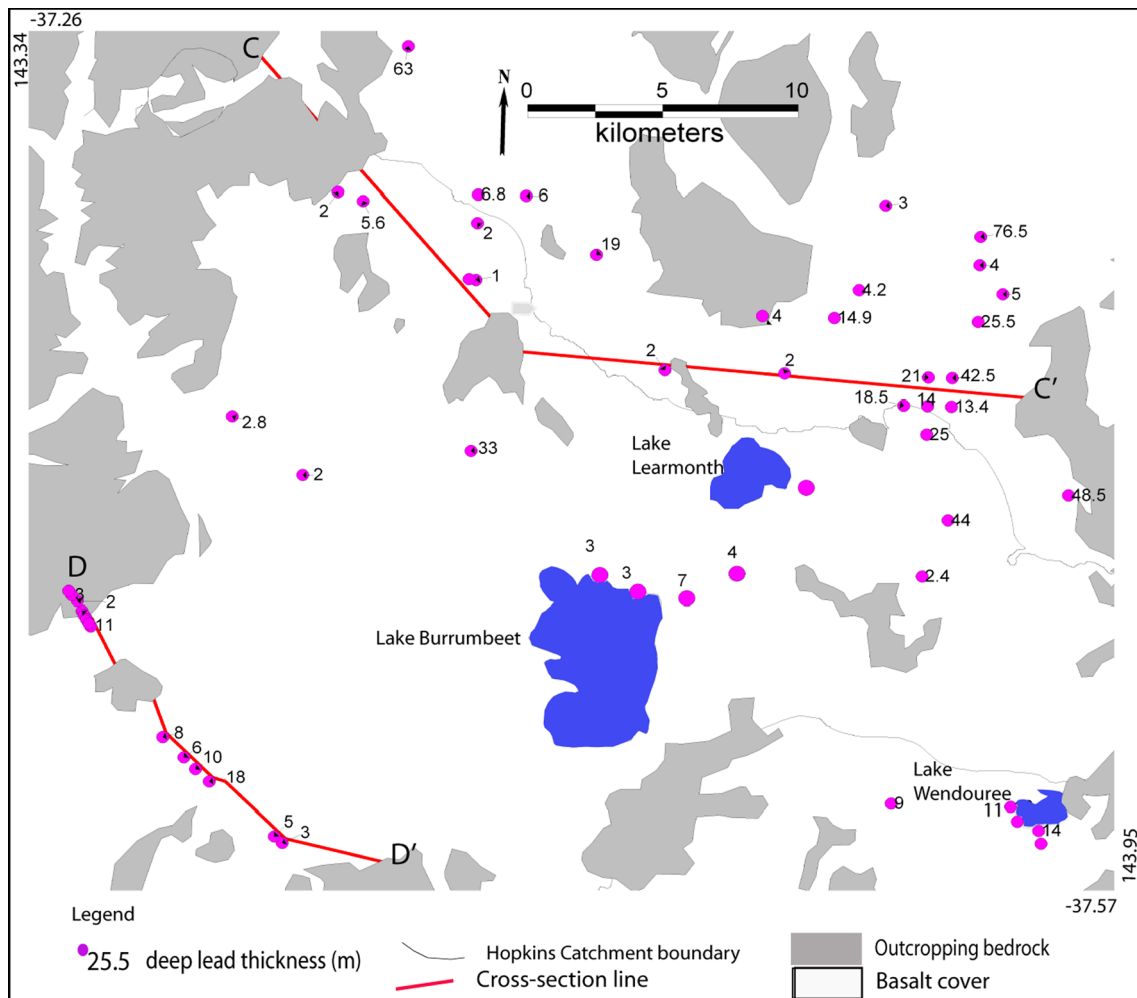
Fig. 10 Deep Lead thickness
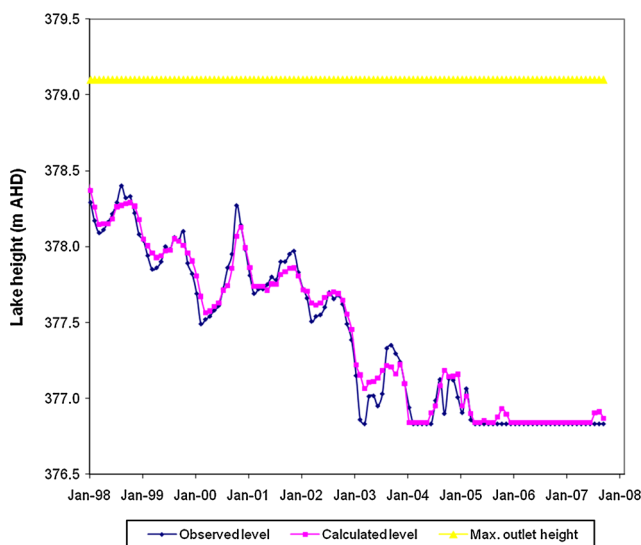
Fig. 11 Lake Burrumbeet measured water levels and comparison to modelled results for the period 1998–2007

The overall water balance suggests that the major influences on lake levels are evaporation (average 39 % of the total water budget), groundwater outflow (average 12 %), direct precipitation (24 %), surface water inflow (20 %) and surface water outflow (3.6 %); groundwater inflow is a minor component (1.3 %). Evaporation has the greatest influence in summer, reaching up to 86 % of the total budget. Direct precipitation and surface runoff dominate in winter (up to 64 and 76 %, respectively); comparison of the long-term monthly components shows that the major influence on the rise in lake level over winter is the increase in stream inflow. Direct precipitation has a comparable influence to surface runoff, due to the relatively large size of the lake (21 km^2) compared to the catchment (298 km^2).

The water balance of Lake Burrumbeet shows that groundwater outflow has a major influence on lake levels (average 31 % of water losses), whereas this has only a minor effect for other lakes in the region such as Lakes

Buninjon and Linlithgow (Yihdego and Webb 2012). The lake water budget confirms that the interaction of Lake Burrumbeet with the surrounding groundwater system is volumetrically significant.

The substantial groundwater outflow for this lake is probably due to leakage through a volcanic neck beneath a maar (Callendar Bay) at one end of the lake into the underlying Deep Lead aquifer. Also there is a likelihood of anomalously high flow through the lake floor where there are outcrops of basalt in the centre of the lake bed. Geological reconstruction of the course of the Deep Lead shows that it runs directly beneath Lake Burrumbeet, and receives leakage from the lake. This loss of groundwater limits the amount of time the lake water is subject to evaporation, and also allows significant salt export, so that the lake generally maintains a low to moderate salinity (except just before it dries out, when it becomes disconnected from the leakage points). Lakes Buninjon and Linlithgow are underlain by thick clay-rich soils, so they do not leak. As a result Lake Burrumbeet is generally much less saline than Lakes Linlithgow and Buninjon (median salinities of 3.7, 17.2 and 12.6 mS/cm, respectively). Lake Buninjon is less saline than Lake Linlithgow, probably because it has a lower area/depth ratio than Lake Linlithgow and dries out less often (Yihdego and Webb 2012).

Geological/hydrogeological synthesis

The water budget results can be incorporated into the hydrogeological framework of the area to give an insight into the dynamics of the lake-groundwater flow system. The lake catchment area is underlain by shallow (basalt) and deep (Deep Lead) aquifers, hydraulically separated by an aquitard, except where it is penetrated by eruption points (Fig. 9). The Deep Lead is predominantly recharged through the eruption points as the case at Lake Burrumbeet, and there is no or only limited inter-aquifer leakage in most areas. The Deep Lead groundwater system flows towards the north; in contrast, flow in the basalt aquifer is towards the southwest.

The Lake Burrumbeet hydrogeological system has a through-flow pattern (Figs. 5, 6), with the groundwater discharge from the lake fed predominantly to the deep aquifer (Deep Lead) via the volcanic neck located in the lake margin; only a relatively small proportion flows into the shallow (basalt) aquifer down-gradient of the lake, because of the low gradient.

Implications

The estimates of groundwater outflow from the lake provide additional information for managing groundwater use and salinity in the area, as the lake is now known to be an

important recharge feature for the Deep Lead aquifer. With regard to salinity, the significant rate of groundwater outflow limits the extent to which evaporation can take place from the surface of the lake, and contributes to the freshness of water that is recharging the Deep Lead aquifer. With regard to groundwater use, this new understanding of the degree of surface-groundwater interaction of Lake Burrumbeet should be considered in any proposed groundwater irrigation schemes near the lake, as this issue was not adequately taken into account when the present project was formulated.

Conclusions

This paper infers leakage of water from Lake Burrumbeet from geological, geophysical and hydrogeological data and a lake water balance analysis for 1998–2008, and accounts for the relative freshness of the lake when compared to the other salt lakes of the region. As a case history, this study provides valuable insights into the nature of groundwater-surface water interactions in lakes associated with lava flows.

In particular, this study has increased the level of understanding of salinity and groundwater flow on the basalt plains in Victoria in Australia, through the development of a hydrogeological conceptual model that considers the lakes to be a surface manifestation of the water table. According to the hydrogeological conceptual model, groundwater outflow from Lake Burrumbeet has a major influence on lake levels (average 31 % of water losses), whereas this has only a minor effect for the other lakes in the basalt plain, because of differences in their hydrogeological setting. The substantial groundwater outflow for this lake is probably due to leakage through a volcanic neck beneath a maar at one end of the lake into the underlying Deep Lead aquifer, which runs directly beneath Lake Burrumbeet. Also there is a likelihood of anomalously high flow through the lake floor where volcanic rocks outcrop in the centre of the lake bed.

Lake Burrumbeet is relatively fresh even though it dries out because it leaks significantly, so there is a significant amount of salt export in the groundwater outflow and the amount of time the lake water is subject to evaporation is limited. This groundwater leakage is sufficiently large to transfer enough salt from the lake to maintain a low to moderate lake salinity. This has a real implication with respect to the development of surface and groundwater salinity at this site.

An understanding of the groundwater flow regime is instrumental in developing a water management strategy for the basin, as there is an ongoing need to efficiently manage the water resources, due to the increasing

population of the urban area on its margin (Ballarat). The information obtained from the present water balance analysis, coupled with the re-construction of the palaeotopography, could be enhanced using additional lines of evidence, including isotopic analysis of groundwater, to constrain the magnitude of output from Lake Burrumbeet, and hence improve the understanding of the role of groundwater in the water budget of the lake and how it is linked to the regional flow system. The long-term research objective would be to conjunctively use isotopic and geochemical techniques, well monitoring networks and remote sensing with numerical flow modelling to describe the spatial and temporal variations in groundwater flow paths, and to define accurately the water table configuration, groundwater recharge, and direction of seepage through the lake bed.

This study presents a case study of how hydrogeological conceptual models and water balance analyses can determine the degree of interaction between a lake and the regional groundwater flow system, and sheds new understanding on controls of lake salinity.

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